VOLUME XLVI NUMBER 2

THE
JOURNAL OF GEOLOGY

February-March 1938

THE EFFECT OF PRESSURE ON THE RIGIDITY OF ROCKS. II

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DISCUSSION

1. PRELIMINARY REMARKS ON AGGREGATES OF CRYSTALS

It is common practice in physics, engineering, and geology to make use of the special case of the theory of elasticity, in which only two parameters are required for the specification of the elastic energy of the body, in dealing with polycrystalline metals and rocks. Strictly, the two-constant theory is valid only for isotropic, homogeneous bodies obeying a generalized Hooke's law. The materials which most nearly satisfy these conditions are probably well-annealed glasses. Single crystals of the cubic system, which embraces most of the common metals, require three parameters for the description of their elastic behavior, while crystals of lower symmetry require from five to twenty-one parameters. The problems encountered in practice generally involve aggregates of crystals, and it is possible in many cases where the crystals are oriented at random to consider the aggregate as isotropic by compensation, if it includes a sufficient number of crystals. Such aggregates are called quasi-isotropic. Under certain conditions the elastic behavior of quasi-isotropic aggregates may be described in terms of two parameters, which should be

derivable from the greater number of parameters of the individual crystals.

The derivation of the elastic parameters of quasi-isotropic aggregates in terms of those of the crystals was treated by W. Voigt[1] in 1887. His theory depends upon the following assumptions: there are no empty spaces, so that the displacements are continuous throughout the aggregate; the individual crystals are large with respect to the radius of action of the molecular forces, but small with respect to the dimensions of the sample of the aggregate under consideration. An additional assumption treated by Voigt as a consequence of the first is that the strains are also continuous throughout the aggregate. The mean stresses are then found by averaging the elastic constants of the individual crystal over all possible orientations. By this process the number of independent elastic paramenters is reduced to two. Evidently the conditions at the boundaries of the crystals are not correctly treated, since continuity of displacement does not imply continuity of strain at the contact of differently oriented crystals. On the other hand, by averaging over a large number of crystals whose symmetry is not too low a nearly correct result may be obtained. An attempt to treat the boundary conditions rigorously has been made by Bruggeman,[2] in whose theory that of Voigt appears as a first approximation, valid for crystals of high symmetry. For our purposes Voigt's simpler theory may as well be retained with the caution that for monoclinic or triclinic crystals it may be appreciably in error, and that, in any case, only approximately correct results are to be expected.

Although especially intended to apply to metal aggregates, for the purpose of discriminating between the multi- and rari-constant theories of elasticity, Voigt's theory could not be applied to these at first, for there were no measurements of the elastic parameters of metal single crystals—the technique of production of such crystals of sufficient size not having been developed until later. The theory was consequently tested with such natural crystals as quartz, fluorite, calcite, and barite, and with aggregates of these crystals occurring

[1] *Lehrbuch der Kristallphysik* (1928 ed.), p. 954.

[2] D. A. G. Bruggeman, *Elastizitätskonstanten von Kristallaggregaten* (The Hague, 1930).

in nature. As we should expect, the values of the moduluses calculated for the aggregates from the measured parameters of the single crystals were considerably larger than those found by direct measurements of the aggregates. Voigt recognized that this difference should be attributed to the porosity of the natural aggregates; he supposed, however, that the ratio E/μ (in which he was specially interested) should be independent of the porosity, if this were regularly distributed. (E is Young's modulus.) Now it is true that, for aggregates with about the same porosity, this ratio is much more nearly constant from sample to sample than either E or μ, but the ratio changes considerably when the pore space of a given sample is greatly altered, as in the present experiments.

Voigt's theory applies only to homogeneous aggregates, that is, to aggregates of a single kind of crystal. Except for porosity, aggregates of metal crystals and of minerals, so long as they are homogeneous, are on the same footing. We should expect a compact aggregate of randomly oriented quartz crystals, for example, to approximate a quasi-isotropic elastic body at least as well as a piece of polycrystalline zinc. In making experiments upon an aggregate, however, the size of the individual crystals must be taken into account; the larger the crystal size, the larger must be the sample upon which the measurements are made. It is difficult to set up any very definite criterion for determining the necessary size of the sample with respect to that of the crystals, but usually it is possible, by working with different samples of a given material, to find a size which will be representative.

When we pass to aggregates of two or more different kinds of crystals we no longer have a well-developed theory as a guide. Treatment of such aggregates as quasi-isotropic bodies depends on an intuition that, if the different kinds of crystal are evenly enough distributed in space and in orientation, there will be no outstanding directional properties for samples of large size. Such an aggregate is treated as "homogeneous" if the composition of samples of a given, arbitrary size is nearly uniform In every case the particular phenomenon which we are considering must be closely scrutinized in order to insure the use of specimens of representative size. In any particular aggregate the question of directional properties may be in-

vestigated experimentally. Special aggregates consisting of two kinds of isotropic particles have been treated by Bruggeman.[3]

A primary cause of confusion in work with rocks is that, while such aggregates may be nearly isotropic and homogeneous, Hooke's law may not hold for the aggregate over the range of strain usually employed, at low pressures. The reason for the failure of this law for many rocks under ordinary conditions is that the areas of contact between the crystals change with the mean applied stress. The condition postulated by Voigt as a basis for his theory—that there be no spaces between the crystals—is not usually satisfied; furthermore, even small stresses are capable, at low pressure, of causing very large changes in the pore spaces.

We can illustrate the effect of high pressure in closing the spaces in aggregates and hence raising their elastic parameters nearer to the theoretical ones. We have values for a very pure marble, a pure fine-grained limestone, and a rather impure quartzitic sandstone (80 per cent quartz). In terms of Voigt's values for calcite and quartz single crystals, the calculated rigidity for compact quasi-isotropic aggregates of these crystals should be $3.73 \cdot 10^{11}$ dynes/cm^2 for the calcite aggregate, and $4.67 \cdot 10^{11}$ for the quartz aggregate. Looking at Table 5 we see that the rigidity of the limestone increases from 1.96 at 1 atmosphere to 2.47 at 4,000, the marble from 1.57 to 3.33, and the quartzite from 3.1 to 4.28 (average), all times 10^{11}. A purer quartzite would have a higher rigidity. Thus the marble and the quartzite show very much closer agreement with the predicted values at high pressure; this improvement is certainly due to the reduction of empty space, and consequently a closer approach to the postulated conditions of the theory. In the case of the limestone, the improvement is much less marked. It is apparently very much more difficult to reduce the pore space in a very fine-grained aggregate than in one of coarser crystals. Voigt's values for the elastic constants of calcite are not perfectly consistent with compressibility determinations by

[3] D. A. G. Bruggeman, "Berechnung verschiedener physikalischer Konstanten von heterogenen Substanzen. III. Die elastischen Konstanten der quasi-isotropen Mischkörper aus isotropen Substanzen," *Ann. der Phys.*, Vol. XXIX (1937), p. 160.

Adams, Williamson, and Johnston[4] and Bridgman;[5] the computed compressibility is about 10 per cent higher than the observed figure.

In passing we should indicate the sort of confirmation which Voigt's theory receives when applied to metal aggregates. Here the question of pore space is much less acute, although differences observed between cast and drawn metals have been attributed to this factor. For a number of metals Voigt's theory gives the relation between aggregate and single-crystal parameters with an error not greater than several per cent; sometimes this error reaches 10 per cent, but this is usually recognized as indicating an appreciable defect of isotropy, or else experimental error. The improved theory of Bruggeman gives notably better agreement in some cases, particularly where the symmetry of the crystals is low. For our purposes such a theory is useful chiefly as a guide in interpreting qualitatively the observed phenomena, for in general the data required for numerical comparisons are lacking.

2. ISOTROPY AND ANISOTROPY

The present method permits the study of the propagation of torsional waves in any specified direction in a mass of rock; we need only to cut a specimen with its axis parallel to the direction in which we are interested. The most obviously anisotropic rock among our samples is the gneiss. Ide's dynamical measurements of Young's modulus[6] E had demonstrated that surprisingly little difference existed in the directions parallel and perpendicular to the foliation. This is confirmed by our measurements of the rigidity; the velocities in the two directions differ by only 3 per cent at 4,000 atmospheres. This is very little greater than the probable error. In view of this it is natural to suppose that most igneous rocks, with no evident bedding or variation of composition, will be very nearly isotropic. The rock

[4] L. H. Adams, E. D. Williamson, and J. Johnston, "The Determination of the Compressibility of Solids at High-Pressure," *Jour. Amer. Chem. Soc.*, Vol. XLI (1918), p. 12.

[5] P. W. Bridgman, "Linear Compressibility of Fourteen Natural Crystals," *Amer. Jour. Sci.*, Vol. X (1925), p. 33.

[6] J. M. Ide, "Comparison of Statically and Dynamically Determined Young's Modulus of Rocks," *Proc. Nat. Acad. Sci.*, Vol. XXII (1936), pp. 81–92; "The Elastic Properties of Rocks: A Correlation of Theory and Experiment," *ibid.*, pp. 482–96.

which showed the greatest anisotropy according to Ide's measurements was the dunite, with differences of E in the hardest and easiest splitting directions amounting to 30 per cent at 1 atmosphere. A difference remains at 4,000 atmospheres, but it is reduced to 2 per cent. Similarly, differences of perhaps 30 per cent among the values for the moduluses of various samples of the Quincy and Rockport granites dwindle to 2 or 3 per cent at 4,000 atmospheres. The Vinal Haven diabase was shown to be very closely isotropic at 1 atmosphere by Ide and also by Zisman.[7] The differences recorded for our specimens are probably due to the fact that the samples 5 and 7 were made from different blocks from the Vinal Haven quarry. The differences for mutually perpendicular colums of the Maryland diabase, of the syenite, of the Mellen gabbro, and of the French Creek gabbro were negligible. In the case of the Sudbury norite the two drill-core specimens came from different borings and apparently differ considerably in composition. Their differences in velocity have no bearing on the question of isotropy, but of course indicate the inhomogeneity of the Sudbury norite body.

Thus it appears that many of the departures from isotropy observed at 1 atmosphere are due to differences of compactness in different directions, which disappear at high pressure. These differences in compactness may be inherent in the material as collected, or produced in the laboratory, as already described. Where they are inherent in the material in the field they may of course affect seismic exploration at shallow depths, but it seems very improbable that they persist much below a kilometer. We cannot exclude the possibility of anisotropy at greater depths due to local conditions of unsymmetrical orientation, stressing, or heating; in view of the results for the gneiss, it appears that such anisotropy would have to assume a very extreme form before it would alter the velocities in different directions by more than a small amount.

[7] W. A. Zisman, "Young's Modulus and Poisson's Ratio with Reference to Geophysical Applications," "Compressibility and Anisotropy of Rocks at and near the Earth's Surface," "Comparison of the Statically and Seismologically Determined Elastic Constants of Rocks," *Proc. Nat. Acad. Sci.*, Vol. XIX (1933), p. 653.

3. LOW-PRESSURE RESULTS

At low pressures the most striking feature is the large increase of velocity produced in many cases by the application of the first few hundreds of atmospheres. This pressure appears to be sufficient to remove most of the looseness of the aggregates; above about 1,000 atmospheres the rate of change of rigidity with pressure begins to be comparable with that found for metals. Except at low pressures the changes are very closely reversible. The deformations of the crystals required for closing up the aggregate at low pressures may be, and often are, inelastic, as a result of concentration of stress upon small regions of contact. At high pressures, in spite of the higher stress at the surface of the specimen, the more uniform distribution of stress over the crystals affords an approach to something nearer a hydrostatic pressure, or, at any rate, to a stress system under which the crystals behave very nearly elastically.

We can be certain that much of the initial looseness of structure of some of the rocks was artificially produced during the course of the experimenting. The loose structure, with low velocity, can be produced by exposing the uncovered rock to high pressure, or by heating at 1 atmosphere. Thus if the covering leaks, or if, as is true in many cases, the specimen is used uncovered for other measurements under pressure, the mechanical properties are sure to deteriorate. Newly prepared specimens of some of the rocks, without this sort of laboratory history, showed no abnormal increase of velocity at low pressure. Furthermore, we are fortunate enough to have duplicate specimens for some of the rocks, which have never been exposed to pressure, and the rigidity of these specimens has been measured at 1 atmosphere by Ide. For these rocks, comparison of our results for 1 atmosphere with Ide's will generally show whether any large effect due to maltreatment has taken place. This comparison can be carried out for all of the granites, the Vinal Haven diabase, the French Creek "norite" (gabbro), the sandstone, gneiss, and marble. In the case of the granite a large part of the looseness was inherent in the unspoiled specimens; the effect of maltreatment was most apparent in the Quincy granite from the 100-foot level, which was the most compact sample of granite, reducing the rigidity in this case from

about $2.7 \cdot 10^{11}$ to about $2.0 \cdot 10^{11}$. Even the value $2.7 \cdot 10^{11}$ is well below the value reached at 500 atmospheres, which is $3.15 \cdot 10^{11}$. The looseness of the French Creek "norite" was apparently inherent in the rock as collected. This is true also of the sandstone. The gneiss was affected somewhat by its treatment, but its initial looseness was very considerable, and the same may be said of the marble.

On the other hand, nearly all of the initial effect observed with the Vinal Haven diabase specimens 3 and 5 was traceable to laboratory treatment. The new specimen 7 showed a nearly linear change of velocity or rigidity with pressure, and its value at 1 atmosphere agreed with the values given by Ide for other fresh specimens. Nearly linear changes were also observed for the Maryland diabase (especially the fresh large specimen 3), the Noranda metadiabase, which came from a depth of 1,700 feet, and the two samples of hypersthenite. A good deal of the initial looseness of the dunite is probably inherent in the rock as collected, but it was considerably increased in certain cases in the laboratory; it is difficult to cover this rock with a tight jacket, and in some cases three or four attempts failed—the jackets leaked before a complete run was secured. The effect of this treatment in lowering the successive values of the initial velocity was easy to observe. In the case of the other rocks we can only guess as to the relative amounts of inherent and induced looseness. It seems very likely that the looseness was mostly induced in the Sudbury norite and inherent in the Mellen gabbro and the syenite.

It is important to emphasize that for rocks in this initial loose condition, where the velocity is so strongly a function of pressure, we must not expect to find the theory of elasticity, as developed for the ideal elastic body, to be applicable. A change of rigidity of the magnitude observed for the granites and other loose rocks means a change of the material in a sense not contemplated by the theory of elasticity, even if this be extended to include finite strain. The same conclusion must be reached on the basis of the work of Zisman on the shortening and thickening of columns of rock under axial stress. Zisman found that the stress was not in general a linear function of the strain; thus the ordinary definition of Young's modulus was not applicable and this modulus had to be specially defined as the slope of the stress-strain curve. This quantity, which we may call a pseudo-

Young's modulus, increased with the mean stress in nearly all cases —by as much as 100 per cent for relatively small loads for certain granite specimens. These results are qualitatively in accord with our observations. They account, as Zisman pointed out, for a large amount of the extravagant disagreement among observed values of various "elastic moduluses" for low stresses.

A recent extension of the theory of elasticity to include finite strains has been published by F. D. Murnaghan.[8] In his theory, which appears to be essentially a more complete and correct description of the geometrical relations between stress and strain, in which the powers of the strain higher than the first are not neglected in the development of the expressions for the stresses, the apparent Young's modulus is found to depend in a natural way upon the strain, even when only two parameters are used in the description of the elastic energy of the body. In this case, however, the apparent Young's modulus decreases as the strain increases; the effect is also incomparably smaller in magnitude than the observed effect in rocks. In other words, an entirely different phenomenon is involved, and we must expect to find that it will often be impossible to describe the elastic behavior of rocks at low pressure in terms of two parameters.

The work of Ide also leads to this conclusion. He showed, in effect, that if we employ the connecting equations of elasticity for an isotropic body, by which any elastic parameter may be computed in terms of two independent ones, then the value obtained for a given parameter, say Poisson's ratio, depends upon which pair of parameters we start with. There is an immediate application of this result to seismological field work. Here we find two principal velocities, which we may take as independent parameters, and if the density is known, we may evaluate the other elastic constants. But if two parameters are not sufficient to describe the behavior of the material under the conditions involved, then the computed values will not agree with those derived from some other type of experiment. Moreover, we should expect the velocities themselves to depend to some extent upon the conditions of the field work, such as

[8] "Finite Deformations of an Elastic Solid," *Amer. Jour. Math.*, Vol. LIX (1937), p. 235.

the amplitude of the motion and the depth in the rock, which is important in the transmission of the energy.

We have at hand good examples of the kind of discrepancy which may arise from this cause. The velocities of artificial explosion waves have been determined by Leet and Ewing,[9] and by Leet[10] in the Quincy and Rockport granites. Computing σ by the usual formula, it is found to be 0.33 at Quincy and 0.31 at Rockport. On the other hand, Ide's measurements on the longitudinal and torsional vibrations of rods and Zisman's on the relative lateral extension of columns under static axial loads agree in giving σ in the neighborhood of 0.1 for low stresses; Adams and Coker, using relatively large stresses, found an average of 0.22 for granite. In other words, since two parameters are inadequate to describe the behavior of this material, whatever third parameter is calculated in terms of the two-constant theory must bear the whole burden of the insufficiency of this theory, and its value will depend upon the choice of independent parameters.

For the present it is not our intention to develop a theory adequate to account for the behavior of loose aggregates. Such a theory would have to take into account the change of areas of contact between crystals with change of applied stress. It seems more useful, except for seismological work at shallow depths, to attempt to discover the conditions under which the elastic behavior of rocks may adequately be treated in terms of the two-constant theory, bearing in mind that at low pressures a considerable range of velocity is possible for the same type of rock, depending upon its history and probably upon the conditions of experimenting. For surface rock of unknown compactness we can merely set maximum values for the velocity; there seems to be almost no lower limit, even for rocks apparently sound.

4. HIGH-PRESSURE RESULTS; CALCULATION OF OTHER PARAMETERS

At high pressures—above about 1,000 kg/cm²—very little trace remains of the possible initial looseness; the subsequent change of

[9] L. Don Leet and W. Maurice Ewing, "Velocity of Elastic Waves in Granite," *Physics*, Vol. II (1932), p. 160.

[10] L. Don Leet, "Velocity of Elastic Waves in Granite and Norite," *Physics*, Vol. IV (1933), p. 375.

velocity is generally small and nearly linear in the pressure. The initial velocity is often a poor indication of the velocity which even a few hundred atmospheres will develop. We interpret this behavior as indicating the more or less complete elimination of the original looseness of structure, which can vary widely among specimens with different geological and laboratory histories. Since a depth of 1 kilometer of rock produces a pressure of several hundred atmospheres, we need not, except for shallow seismic exploration, concern ourselves with this complicated initial condition, but may focus our attention upon the much simpler and less ambiguous results obtained at high pressure.

From Table 6, which gives the pressure coefficients of velocity and of rigidity, it appears that in many cases these coefficients are smaller in the pressure range 3,000–4,000 than in the interval 2,000–3,000; the most reasonable explanation of this effect seems to be that in these cases a certain amount of "porosity" or looseness, in the sense of small lenticular intercrystalline spaces, still remains. As the exhaustion of the empty space proceeds, much greater increments of pressure are required to produce a given relative reduction of the remaining empty space. In nonporous, polycrystalline metal aggregates, the increase of rigidity with pressure is very closely linear. For the perfectly compact mineral aggregate a similar effect would be expected, and it is approximately realized for a number of our samples, even with the present pressure range. The extension of the pressure range would no doubt bring greater uniformity into these coefficients. Certain differences would probably remain, however, for the coefficients are probably more sensitive to structure than are the velocities themselves.

In most cases the pressure coefficient of rigidity for the 3,000–4,000 interval is already of the same order of magnitude as the coefficient for metals. These range from about $1.1 \cdot 10^{-6}$ for palladium[11] to $7.6 \cdot 10^{-6}$ for aluminum,[12] while for the rocks the range is from $1.5 \cdot 10^{-6}$ to $21.2 \cdot 10^{-6}$.

[11] P. W. Bridgman, "The Effect of Pressure on the Rigidity of Several Metals," *Proc. Amer. Acad. Arts and Sci.*, Vol. LXIV (1929), p. 39.

[12] Francis Birch, "The Effect of Pressure on the Modulus of Rigidity of Several Metals and Glasses," *Jour. App. Phys.*, Vol. VIII (1937), p. 129.

The pressure coefficient of rigidity for an isotropic solid, in which the energy of deformation depends upon only two parameters, say λ and μ, may be derived from Murnaghan's extension of elastic theory, already cited. We suppose that λ and μ are independent of the pressure, or of stress in general. Nevertheless, this theory gives an increase of apparent rigidity under hydrostatic pressure, expressed by

$$\frac{1}{\mu'}\frac{\partial \mu'}{\partial p} = \beta\left(\frac{\lambda}{\mu} + 3\right)$$

where β is the volume compressibility and μ' the apparent rigidity, which is what we in fact measure. The derivation of this from Murnaghan's theory will be given in another place. The usual elastic theory, valid for infinitesimal strains, predicts no such change, for here the stresses are linear functions of the strains, and stresses due to different simultaneous systems of strains are simply added. The interesting point about Murnaghan's theory is that it permits surprisingly correct values for the pressure coefficients of rigidity of metals to be derived from the measured values of compressibility and elastic constants. We shall see that the same is true for rocks at high pressure.

The compressibilities of the samples which we have used have been measured in this laboratory, and the results will be given in another paper. For the moment we shall simply take the values in which we are interested, the compressibility at 4,000 kg/cm^2; these will be found to agree in general with the values for similar (in several cases, identical) rock types given by L. H. Adams and his collaborators. These compressibilities, and various derived quantities, will be found in Table 10. All are for 4,000 kg/cm^2 and 30° C. The ratio λ/μ is given approximately by $2\sigma/(1-2\sigma)$. Then we find for the pressure coefficient of rigidity at 4,000 kg/cm^2 the following values, which should all be multiplied by 10^{-6}: Quincy granite, 7.25; syenite, 7.12; Maryland diabase, 4.88; hypersthenite, 3.62; dunite, 3.3. If these are compared with the measured values in Table 6 we find that the only really large departure from the mean measured values is for the dunite; the measured values in this case are considerably too large except for one specimen. This agrees with our general impression that an unusual amount of porosity continues to ex-

ist in this rock. In comparing these computed and observed values the rather large experimental uncertainty should be taken into account.

This comparison is of special importance for the support which it gives to the belief that at high pressures we can compute the remaining, unmeasured elastic parameters of rocks in terms of the two measured ones, rigidity and compressibility; it suggests very strongly that two independent parameters are sufficient to describe the elastic behavior of these rocks. A complete demonstration that the two-constant theory is applicable has been carried out for a single rock, Vinal Haven diabase, by Ide and by Zisman. Several others of the present specimens, not available for Ide's work, would have served equally well—for example, the Maryland diabase or the hypersthenite. These are the rocks of greatest initial compactness. But under pressures of the order of 1,000 atmospheres, all of the rocks are probably so compact that the two-constant theory may be applied with good precision.

Consequently in Table 10 we have computed Young's modulus, Poisson's ratio, and the velocity of longitudinal waves in the infinite medium, in terms of the measured rigidity and compressibility. Examination of these results will be deferred while we consider another point.

At high pressure we find that the rocks fall into the following order of decreasing torsional velocity: dunite and pyroxenites, diabases and gabbros, quartzite, granites and granite gneiss, marble, syenite, limestone. This is very nearly the order found for the bulk modulus by Adams.[13] There are, however, a few striking differences: dunite and pyroxenite are indistinguishable; quartzite, instead of being at the bottom of the list, where its high compressibility would place it, is well up with the gabbroic rocks; syenite, less compressible than the

[13] L. H. Adams and E. G. Williamson, "The Compressibility of Minerals and Rocks at High Pressures," *Jour. Franklin Inst.*, Vol. CXCV (1923), pp. 475–529; L. H. Adams and R. E. Gibson, "The Elastic Properties of Certain Basic Rocks and of Their Constituent Minerals," *Proc. Nat. Acad. Sci.*, Vol. XV (1929), p. 713; L. H. Adams and R. E. Gibson, "The Compressibilities of Dunite and of Basalt Glass and Their Bearings on the Composition of the Earth," *ibid.*, Vol. XII (1926), p. 275; L. H. Adams, "The Compressibility of Fayalite and the Velocity of Elastic Waves in Peridotite with Different Iron-Magnesium Ratios," *Gerlands Beitr. zur Geophys.*, Vol. XXXI (1931), p. 315.

granites, nevertheless has a markedly lower velocity. Purely qualitatively, direct measurements of velocity give noticeably different relationships between the different rock types than do compressibility measurements alone.

TABLE 10

ELASTIC PARAMETERS OF CERTAIN ROCKS AT 4,000 KG/CM² AND 30° C.

Rock	E Dynes/Cm²	μ Dynes/Cm²	β Cm²/Dyne	σ	V_p Km/Sec	V_s Km/Sec
Quartzitic sandstone	$9.6 \cdot 10^{11}$	$4.28 \cdot 10^{11}$	$(24.0) \cdot 10^{-13}$	0.118	6.08	4.00
Solenhofen limestone	6.3	2.47	21.4	.276	5.54	3.08
Marble, Vermont	8.7	3.33	(13.9)	.299	6.51	3.49
Granite:						
Quincy 1	8.49	3.45	19.2	.229	6.08	3.61
Rockport	8.36	3.36	18.5	.243	6.24	3.59
Syenite, Ontario	8.04	3.15	16.9	.274	6.05	3.36
Norite, Sudbury 2	9.67	3.81	14.4	.268	6.49	3.65
Diabase:						
Vinal Haven	11.40	4.46	11.7	.277	6.97	3.88
Maryland	11.34	4.42	11.6	.281	6.96	3.83
Gabbro:						
Mellen	10.43	4.00	11.4	.302	6.96	3.71
French Creek	12.20	4.80	11.3	.270	7.15	3.98
Pyroxenite:						
Hypersthenite	16.90	6.86	9.6	.230	7.83	4.58
Bronzitite	17.00	6.80	8.9	.249	7.86	4.55
Dunite	17.30	6.84	8.3	0.262	8.05	4.57

The clue to these differences is indicated in Table 10. Wave velocities can be calculated from compressibilities only with the aid of an assumption as to one other parameter. Adams computed his wave velocities throughout on the assumption that $\sigma = 0.27$, a value found by seismologists for moderate depths. If σ remains constant the velocities are proportional to the square root of the bulk modulus. But we know that for any aggregate containing an appreciable amount of quartz σ must be smaller than 0.27; if we use Voigt's values and theory we find $\sigma = 0.07$ for a pure quartz aggregate. In

Table 10 we see that for our impure quartzite at 4,000 kg/cm², σ = 0.118. The effect of lowering σ is to give higher velocities for a given compressibility. Thus the granites, where σ is about 0.23 or 0.24, have higher velocities than were derived by Adams, since the effect of the quartz in reducing σ is more important than its effect in reducing the bulk modulus.

The values of compressibility in parentheses were not measured on our specimens; the compressibility of the quartzite was estimated in terms of its composition, and the compressibility of the marble is the value given by Adams and Williamson. There is reason to believe that the latter value is too low for this marble at 4,000 atmospheres; a higher value would give a lower value of σ and lower values of E and V_p.

For all the quartz-free rocks except the pyroxenites we find that σ is indeed in the neighborhood of 0.27; for such rocks we therefore expect to be in agreement with the velocities given by Adams, Williamson, and Gibson, and this is the case. Our values for syenite, average diabase or gabbro, and dunite, obtained by the use of high-frequency vibrations, are within a few per cent of the values obtained from static compressibilities with the assumption that σ = 0.27.

The values of σ given in Table 10 are all greater than those obtained at 1 atmosphere by Zisman and by Ide; the difference varies from about 3 per cent for the Vinal Haven diabase to over 100 per cent for the granites. This change is evidently another indication of the initial degree of compactness of the rock.

The closeness of the velocities in pyroxenite to those in dunite, in spite of the higher compressibility of pyroxenite, is again to be attributed to lower values of σ. The value of σ for a homogeneous aggregate can, as we pointed out above, be deduced from the elastic coefficients of the single crystal, and the known values for quartz permit us to account for the peculiar properties of the quartz aggregate. No values are available for single crystals of any pyroxene; they would be of interest in connection with these unexpected results for the aggregate. We are not yet able to compute the elastic coefficients for such crystals, or their interrelations, in terms of the physical structure. Of course, the value 0.27 requires theoretical explanation just as urgently as does 0.24.

5. DAMPING AND DISPERSION

Since we are primarily interested in the velocity of propagation of waves produced by earthquakes and by artificial earth shocks, it is necessary to consider how the velocity of such waves, whose periods are generally not less than 0.1 second, may differ from the velocities which we have studied, where the periods are from five hundred to three thousand times smaller than this. With the present method it is relatively easy to go to higher frequencies, by using higher harmonics, but lower frequencies can be obtained only by using longer specimens or by loading. Specimens long enough to reduce the frequency by a factor of several hundred are clearly not practicable, and we have not yet attempted to use heavily loaded specimens. Thus for the moment our conclusions as to the effect of frequency upon velocity must rest upon indirect evidence.

This is of several varieties, all of which indicate that the effect is probably small. In the first place, there appears to be no change of importance when the frequency is increased. For longitudinal vibrations of rods there is a change of velocity with frequency which depends upon the ratio of diameter to length of the rod. Ide has shown, for a number of different rocks, that even at 1 atmosphere, and for such coarsely crystallized material as granite, the sequence of overtones up to the eighth harmonic is given with fair precision by the theory developed for perfectly elastic solids. This indicates that there is no abnormal dependence of velocity upon frequency in rocks at high frequencies. The same result is obtained when we study the overtones of the torsional frequency, although we have not obtained so many harmonics. It would seem reasonable to expect that, if anomalous velocities were to be found, this would be for very high frequencies, where the wave-length begins to approach the crystal diameter, but no such effect is observed with the present frequency range, where, however, the shortest wave-length ever used (by Ide) is still about ten times the mean crystal size. The fundamental wave-length in our work, which is twice the length of the specimen, is always from one hundred to several thousand times the average crystal diameter. At lower frequencies the effects of crystal size would be still less important.

For the velocities at low frequencies we are forced to appeal to field observations of artificial explosion velocities. This procedure is not perfectly satisfactory, for we have already seen how seriously the velocities at shallow depths may depend upon the condition of the rock. Also, the field velocities are "group velocities," for an undefined group of waves of different frequencies, which are all, however, relatively low. We have the results of Leet and Ewing and of Leet for three quite definite formations—the Quincy granite, Rockport granite, and Sudbury norite—from which we have also a few good samples. The best comparison to make is with the large cylinders whose velocities are given by Ide, since these have not been exposed artificially to pressure and represent most closely the actual condition of the surface rock. Ide's average torsional velocity for the Quincy surface granite (4 samples) is 2.67 km/sec; for the Rockport granite (2 samples), 2.66; for the Sudbury norite (2 samples), 3.57. The field measurements of this velocity gave, in the same order, 2.48, 2.70, and 3.49 km/sec. Thus in one case the field velocity was higher, in two cases slightly lower; the agreement is surprisingly good, in view of the small number of samples and the possible variations at low pressure. There is no evidence here that the slower vibrations produced in the field traveled with a velocity systematically different from that of the very rapid vibrations used in the laboratory.

As shown in Table 8, the damping in most of the rocks, at the resonant frequencies of these experiments, is somewhat greater than in glass at ordinary temperature and, in general, is less at high pressure than at low. The greatest changes with pressure are exhibited by the rocks with the largest crystals; the fine-grained rocks, though in some cases among the best vibrators at one atmosphere, do not usually show so much improvement under pressure. The amount of damping indicated in Table 8 is in no case sufficient to influence perceptibly the velocity of the waves for which these damping coefficients were determined. The damping at one atmosphere was often too great to permit even a rough measurement and conceivably affects the velocity to a small extent. A clear presentation of the theory underlying the specification of the damping in terms of the "con-

stant," $Q(=f_0/\Delta f_0)$, and the relation of this coefficient to the logarithmic decrement of free vibrations is given in recent papers by C. Zener.[14]

If we attempt to find what amount of dispersion may be attributed to the effect of damping, we are faced with the fact that we are ignorant of the way in which damping depends upon the frequency, especially at low frequencies. Although formidable speculative treatments of this problem have been proposed, the experimental evidence available, even for metals and glasses, is meager. The recent work of Wegel and Walther[15] suggests that to the best first approximation the energy loss per cycle is independent of frequency over a wide frequency range (1,000–100,000 cycles per second). A similar conclusion results from the work of A. Gemant and W. Jackson[16] on glasses at low frequencies (0.1–10 cycles per second), and the loss per cycle for glasses in these widely different frequency ranges is very nearly a constant. L. Boltzmann[17] long ago observed this effect for slow torsional vibrations of metal wires, and gave a general theory in terms of "elastic afterworking"; this work has recently been extended and applied to certain geophysical questions by B. Derjaguine.[18] The theory leads to the prediction of some dispersion for very low frequencies, but it appears that the frequencies required to give a measurable change of velocity are lower than those effective in the transmission of seismic waves; for higher frequencies this theory conforms with the generally observed result of a loss per cycle independent of frequency, and negligible dispersion. The same

[14] "Internal Friction in Solids. I. Theory of Internal Friction in Reeds," *Phys. Rev.*, Vol. LII (1937), p. 230; "Internal Friction in Solids. II. General Theory of Thermoelastic Internal Friction," *ibid.*, Vol. LIII (1938), p. 90; C. Zener, W. Otis, and R. Nuckolls, "Internal Friction in Solids. III. Experimental Demonstration of Thermoelastic Internal Friction," *ibid.*, Vol. LIII (1938), p. 100.

[15] R. L. Wegel and H. Walther, "Internal Dissipation in Solids for Small Cyclic Strains," *Physics*, Vol. VI (1935), p. 141.

[16] "Internal Friction in Solid Dielectrics," *Phil. Mag.*, Vol. XXIII (1937), p. 960.

[17] *Wiss. Abhand.*, Vol. I, p. 616.

[18] "Propagation des ondes élastiques dans un milieu non idéalement élastique," *Gerlands Beitr. zur Angew. Geophys.*, Vol. IV (1934), p. 452.

conclusion may be reached on the basis of an elastic afterworking theory proposed by H. Jeffreys.[19]

Although Boltzmann's theory appears to account roughly for the experimental results, it is rarely rigorously applicable, and various other ways of accounting for the internal damping of solids have been proposed. We may briefly consider the consequences which are readily deduced from two of these alternative treatments. Voigt and others have supposed that the damping in solids depends upon a mechanism resembling viscous flow in liquids; Jeffreys[20] has named this condition "firmoviscosity." In this case the velocity of propagation of a plane wave will indeed depend upon the frequency, but in such a way that departures from a nearly constant value will be apparent only at very high frequencies. If this effect existed it should be observable at the higher harmonics, but it has not been found for rocks even when, as in Ide's work, the frequency was increased eightfold. The energy loss per cycle in this case should be proportional to the frequency. There seems to be little evidence that this law is ever applicable.

Another way of treating the damping is in accordance with Maxwell's relation,[21] called by Jeffreys the "elasticoviscous" law. According to this the velocity would be lowest for the lowest frequencies, increasing and approaching a limit as the frequency increased. But if we evaluate the "relaxation time" which enters into this law, for granite at low pressures, we find about 10^{-3} second, with the result that the velocity for a frequency of 1 cycle per second would be only about one-tenth of the velocity at high frequency. This is scarcely compatible with the small difference between Leet's velocities and those found by Ide and by ourselves. In the case of metals, slow vibrations with the torsion pendulum give very closely the same rigidities as do the very rapid vibrations of short cylinders. There seems, finally, to be no evidence that any marked dependence of velocity upon frequency exists for rocks, although we have not,

[19] "On Plasticity and Creep in Solids," *Roy. Soc. Proc.*, Vol. CXXXVIII (1932), p. 283.

[20] H. Jeffreys, *The Earth* (2d ed.; Cambridge University Press, 1929), p. 264.

[21] J. C. Maxwell, *Phil. Mag.*, Vol. XXXV[4] (1868), p. 134.

as yet, any experiments definitely bearing upon this question over a sufficiently great range of frequencies. A small amount of dispersion, giving extreme velocity differences of several per cent, is not inconceivable, especially at low pressure with the looser rocks. The effect of temperature upon the damping in rocks is not perceptible with our present small temperature range.

6. EFFECT OF TEMPERATURE; EFFECT OF COMBINED TEMPERATURE AND PRESSURE IN THE CRUST

In Tables 6 and 7 we have given values for the temperature coefficient of velocity at 4,000 atmospheres, between 30° and 100° C., and for the pressure coefficient of velocity at 30°, for intervals of 1,000 kg/cm² between 1,000 and 4,000. One of the temperature runs (Quincy granite 5) was made at 1,000 kg/cm²; at these pressures the effect of pressure on the temperature coefficient of velocity appears to be small, although there is possibly a large change in this coefficient between 1 and 1,000 atmospheres. Experience has shown that true temperature coefficients are difficult to obtain with rocks at 1 atmosphere, since large irreversible changes, always corresponding to "permanent" *decreases* of velocity, are produced by heating and cooling, although with the present small temperature range this might not be important. Under our conditions, also, there are irreversible changes, but they are usually small. Thus in Table 7 the second column shows the "permanent" *increase* of velocity observed after heating from 30° to 100° and returning to 30°, all at 4,000 atmospheres. Such changes are not large enough to affect the temperature coefficients seriously, and these runs may be considered very nearly perfectly reversible. Two other runs, not listed in the table, with Quincy granite 1, gave "permanent" increases of 0.64 and 0.95 per cent; the average temperature changes were about the same as for the two granites with very small irreversible effects. These were the largest irreversible effects of temperature observed in this work, but it is not impossible that a higher temperature range would produce still greater effects. We may observe that, for rocks under these conditions, the result of these irreversible changes is always to *increase* the velocity at a given pressure.

Taking the irreversible changes into account, the probable error

in the temperature coefficient in the case of any particular rock, except the quartzitic sandstone, is not greater than about 10 per cent. The change for the sandstone is so small, and the permanent change relatively so large, that the precision is low; the outer limits for the coefficient are perhaps -4.10^{-6} and -13.10^{-6}, which are still very much smaller than for any of the other rocks. The coefficients have been corrected for change of length with temperature; fortunately this correction is not important in most cases, for the thermal expansions, even at 1 atmosphere, are not too well known. We have assumed that the linear thermal expansion at 4,000 atmospheres is not notably different from its value at 1 atmosphere.

While there seems to be little doubt as to the effect of temperature upon the granites, the values for the other rocks are more scattered, with, on the whole, an increasingly negative temperature coefficient for the more basic rocks. Certain rocks appear to be exceptional: the Sudbury norite, the Mellen gabbro, and Sample 1 of dunite. This specimen of dunite was the smallest of the four, in length and in diameter, and we therefore consider it the least representative. The diabases and the pyroxenites agree fairly well among themselves. It is certainly desirable to find this coefficient for a greater number of samples and, of course, over a greater temperature range. The *increase* of velocity with *increasing* temperature for pyrex glass and for obsidian should be noticed; this effect has already been observed for these glasses, and for silica glass, at 1 atmosphere.[22]

A precise determination of the effect of pressure and temperature in suitable combination upon the velocities, combined with correspondingly precise seismological determinations of velocity as a function of depth in the upper crust, would enable us, not only, in all probability, to identify the material at different depths, but also to estimate the thermal gradient at these depths. Neither the laboratory nor the seismological results are sufficiently precise or extended to allow us to get very far with these problems for the moment. It is not without interest, however, to push the present information to the limit, partly as example and partly for the suggestive results obtainable.

[22] J. M. Ide, "The Velocity of Sound in Rocks and Glasses as a Function of Temperature," *Jour. Geol.*, Vol. XLV (1937), pp. 689–716.

The velocity gradient in a given layer may be expressed in terms of the pressure and temperature gradients of velocity, and the depth gradients of pressure and temperature in the layer, which is supposed to be chemically homogeneous. Thus if we take the depth equal to z with the positive axis downward, we have $\Delta v/\Delta z = (\Delta v/\Delta p)_t(\Delta p/\Delta z) + (\Delta v/\Delta t)_p(\Delta t/\Delta z)$. The variation of velocity with depth is not known for the shallow layers. By putting for $\Delta v/\Delta z$ the limiting value for the upward refraction of waves, we may find the maximum permissible temperature gradient in any layer consistent with this assumption—that the waves are not refracted downward; thus, that no "shadow zones" exist. The limiting condition is expressed by $dv/dr_s = \bar{v}/r$, where $\bar{v}$ is the apparent surface velocity, r the radius, and r_s the radius at the point of greatest penetration. For our purposes the difference between r and r_s may be neglected, and r taken as 6300 km; the difference between $\bar{v}$ and v is also insignificant. We identify $\Delta v/\Delta z$ with $-(dv/dr)$, and the foregoing relation becomes an equation for $\Delta t/\Delta z$, giving $\Delta t/\Delta z = [-v/r - (\Delta v/\Delta p)_t(\Delta p/\Delta z)] \div (\Delta v/\Delta t)_p$.

TABLE 11

AVERAGE VALUES OF PRESSURE AND TEMPERATURE COEFFICIENTS OF TORSIONAL VELOCITY

ROCK	$\frac{1}{v}\left(\frac{\Delta v}{\Delta p}\right)_t \cdot 10^6$			$-\frac{1}{v}\left(\frac{\Delta v}{\Delta t}\right)_p \cdot 10^6$	$\frac{\Delta p}{\Delta z}$ KG/CM² PER KM
	1,000–2,000	2,000–3,000	3,000–4,000		
Granite	14	6	3	40	270
Diabase	.....	.....	1.8	62	300
Pyroxenite	.....	.....	1.6	90	330
Dunite	.....	.....	3?	140	330

The average values which appear to be of the greatest interest are given in Table 11. If the effect of pressure on $(\Delta v/\Delta t)_p$ is small, so also will be the effect of temperature on $(\Delta v/\Delta p)_t$. The effect of pressure on $(\Delta v/\Delta p)_t$ is certainly to decrease this coefficient, and the effect of temperature on $(\Delta v/\Delta t)_p$ is very probably to give a greater negative coefficient, although this may be offset to some extent by the increase of velocity due to annealing. Thus the prob-

able variations of these coefficients are such that the values which we obtain for the maximum temperature gradient are probably too high. These values are given in Table 12.

In terms of our assumptions we conclude that if the thermal gradients are less than these, waves may be refracted upward; if they are greater, then downward refraction occurs. The values for the granite and diabase are really the only ones of interest, since there is no reason to suppose that pyroxenite or dunite forms layers so near the surface, and the coefficients may change appreciably for

TABLE 12

MAXIMUM THERMAL GRADIENTS AT VARIOUS DEPTHS, CONSISTENT WITH UPWARD REFRACTION OF WAVES

Rock	Pressure Range, Kg/Cm²	Approximate Depth, Km	$\frac{\Delta t}{\Delta z}$ (°C./Km)
Granite	1,000–2,000	3.7 – 7.4	98.5
	2,000–3,000	7.4–11.1	44.5
	3,000–4,000	11.1–14.8	24.2
Diabase	3,000–4,000	11–15	11.3
Pyroxenite	3,000–4,000	11–15	7.7
Dunite	3,000–4,000	11–15	8.2?

the conditions under which these materials may be expected to exist. The conditions of actual existence of the granite are, however, reasonably well covered by the experiments, and the results suggest that the maximum permissible temperature gradient for upward refraction is probably greater than the actual gradient throughout this layer. On the other hand, a layer of diabasic material at a depth of 15 or 20 km might very well give downward refraction; in all likelihood the existence of "shadow zones" such as would be produced by a decrease of velocity with increasing depth in a thin layer cannot be altogether excluded by present seismological evidence. The existence of such zones would imply either greater thermal gradients in some region than we have indicated or a greater decrease of velocity with increasing temperature at higher temperatures or, of course, different or gradually changing materials. These

alternatives are all conceivable, so that the existence of shadow zones seems not impossible. The discovery and delimitation of such zones, or their exclusion by close networks of observing stations, would very considerably augment our precision in estimating the composition of the crust and its temperature.

7. COMPARISON WITH SEISMIC VELOCITIES IN SURFACE LAYERS

The principal object of this work is to discover what values the seismic velocities would have in large layers of the particular rocks which we have studied, if they were to exist at various depths in the

TABLE 13

TEMPERATURES, PRESSURES, AND VELOCITIES IN SURFACE LAYERS

DEPTH KM.	TEMPERATURE, °C.	PRESSURE, KG/CM²	KM/SEC		MEAN TEMPERATURE °C.	MEAN PRESSURE KG/CM²
			V_p	V_s		
0.....	10	1				
			5.0	3.3	55	500
4.....	100	1,000				
			5.5	3.36	290	3,000
17.....	480	5,000				
			6.3	3.65	720	8,500
42.....	960	12,000				
			7.8	4.35	.....	.....

crust. A complete solution of this problem requires not only an increased range of laboratory conditions, but also a better knowledge of the temperatures at the relevant depths; the pressures are well enough known. As suggested in the preceding section, laboratory data, combined with precise seismological observation and analysis, may be capable of increasing our insight into the problem of the temperature.

For the moment, in order to make use of such material as we now possess, we can adopt certain estimates as to temperature and stratification of the crust; we have assembled in Table 13 a combination of temperature and velocity data, as function of depth, taken

with a few minor changes from A. Holmes[23] and from E. Tillotson.[24] These data are probably not mutually consistent, but for our purpose this is not of great importance. Rough values of pressure are also given.

Of the uppermost layer, consisting chiefly of sedimentary or metamorphic rocks, we have nothing to say at present. The next layer corresponds to the "granitic" layer of Jeffreys and others; it is the layer, aside from the one above it, whose conditions are most completely reproduced in the present work. To compute the velocity under the mean pressure and temperature estimated for this layer, we have to extrapolate from 100° to 290° C.; no extrapolation with regard to the pressure is required. Using the average temperature coefficient found at high pressure for the granite specimens of this paper we find that the velocity observed at 30° and 3,000 kg/cm² should be reduced by about 1 per cent at 290°; thus for our average granite the S velocity would still be 3.53 km/sec under the mean conditions, or nearly 5 per cent too high. The average temperature coefficient required to account for this difference would be six times as great as the observed one. The Westerly granite, on the other hand, gives an estimated velocity at 290° and 3,000 kg/cm² only 2 per cent too high; thus these seismic velocities are perhaps consistent with a granite of relatively low quartz content. If we take Jeffreys' value for V_s in this layer, 3.3 km/sec, the discrepancies are considerably greater, reaching nearly 4 per cent even for the Westerly granite. Our sample of gneiss gives very closely the same velocity as the average granite.

Various possible interpretations of this difference suggest themselves. It is possible that the present granite samples do not represent all "granitic" rocks and that more variation exists among rocks which would be so designated than is indicated by our closely grouped results. It is possible that the path of waves traveling in the "granitic" layer includes an appreciable amount of rock in which the

[23] "Radioactivity and the Earth's Thermal History," *Geol. Mag.*, Vol. LII (1915), p. 11.

[24] "On an Earthquake Near Imotski, Yugoslavia, 1923, March 15," *Mon. Not. Roy. Astr. Soc. London, Geophys. Supp. 2* (1931), p. 416.

velocities are lower, such as metamorphic rock; as against this must be balanced the likelihood of large injections of rock with higher velocities from the deeper layers. Finally, there is the possibility of systematic error in the seismic velocities. The questions of the present experimental error and of dispersion have already been considered.

The low velocities which are found in the laboratory for granites at low pressure may be observed in exposed bodies of granite, over short distances; as the distance increases, however, the average velocity should rapidly increase. If a layer of some other material exerts a sufficient pressure on the upper surface of the granite, then this region of rapid increase will not exist in the granite, though it may now be found in the overlying layer. The travel-time curve which would be given by an exposed granite layer with the depth-velocity relation indicated by the results of this paper will soon be worked out, and this will permit a closer comparison with the seismological data.

There is some evidence, however, that these seismic values for the "granitic" layer are too low. Leet[25] has given the velocities for the P and S waves as 6.0 and 3.5 km/sec, respectively, for the upper layer in New England; these were determined from timed quarry blasts at well-determined positions. He has also cited the experiments of Wiechert near Göttingen, where a longitudinal velocity of 6.0 km/sec was found at depths from 2 to 8 km, and the work of Wood and Richter[26] in California, who found about the same velocity. In this latter work a very much smaller P velocity (5.58 km/sec) was obtained when the known origin time was ignored and the velocity computed from the difference in arrival times at two stations. The consistency of these higher values, determined without many of the assumptions necessarily involved in the reduction of earthquake records, with our own values for granite, tends to con-

[25] L. D. Leet, "The Provincetown, Massachusetts, Earthquake of April 23, 1935, and Data for Investigating New England's Seismicity," *Proc. Nat. Acad. Sci.*, Vol. XXI (1935), p. 308; "Seismological Data on Surface Layers in New England," *Bull. Seism. Soc. Amer.*, Vol. XXVI (1936), p. 129.

[26] H. O. Wood and C. F. Richter, "A Second Study of Blasting Recorded in Southern California," *Bull. Seism. Soc. Amer.*, Vol. XXIII (1933), p. 95.

firm a belief in the existence of granitic layers in these widely separated regions, and also in a possible defect in the technique of finding velocities in the upper layers from near-earthquake data. Until these discrepancies are cleared away by the seismologists, little progress can be made toward more definite identification of the materials of these layers.

Consideration of the "intermediate" layer is affected by the same uncertainties, with the additional ones due to the greater uncertainty as to the actual temperatures and pressures and to the much longer extrapolation required from the realized laboratory conditions. The discussion of this layer is also hampered by the fact that we have no samples of such rocks as granodiorites or diorites, which have sometimes been suggested for this region. We may, however, estimate the velocity in our "average" gabbroic or diabasic material under the mean conditions given in Table 13 of 720° C. and 8,500 kg/cm². We find that the pressure would increase the velocity by not more than 0.8 per cent, and the temperature would decrease it by probably not less than 4.3 per cent; the net decrease would thus be not less than 3.5 per cent, which would bring the average torsional velocity down to 3.73 or less. In view of the greater extrapolation and the probable effect of higher temperature in producing a greater than linear decrease, this value is not seriously inconsistent with the observed value 3.65 (which is often given as 3.7). It is certainly not possible to say, however, that no other material would satisfy the requirements. Variations among the diabasic or gabbroic rocks are, furthermore, sufficiently great to allow some choice.

A narrower range of choices will perhaps remain when the effect of temperature upon the *P*-velocity has been determined; we have at present no indication that any general relation exists between the temperature coefficients of the *P* and *S* velocities. If these velocities depend upon two independent parameters their temperature variations may also be independent. A more serious difficulty in the identification of these layers is the likelihood of inhomogeneity, both in depth and in the other dimensions. If this is very considerable, however, it should be eventually discovered from the seismograms.

A relatively simple situation exists with regard to the lower layer,

for we have only two types of rock with velocities approaching the required ones, given in Table 13 as 7.8 and 4.35 km/sec. The extrapolation with regard to temperature is so great that our discussion becomes extremely speculative; the pressure, 12,000 kg/cm², although greater than our present range, is nevertheless reached in the experiments of L. H. Adams, and we know that at ordinary temperature, at least, no sudden change takes place in these rocks. Extrapolation from 4,000 to 12,000 kg/cm² at ordinary temperature thus presents comparatively little risk; the increase of V_s is not more than 1.3 per cent for the pyroxenite or 2.4 per cent for the dunite; for the latter we have used the coefficient $3 \cdot 10^{-6}$, which is probably greater than the average for this range. Straight-line extrapolation for the effect of temperature gives a decrease of 8.4 per cent for the pyroxenite and 13 per cent for the dunite; the net decreases are 7.1 per cent and 10.6 per cent, giving the velocities 4.25 and 4.09 km/sec for the pyroxenite and dunite, respectively. It is not easy to see how the velocities can be much greater than these unless the pressure has a very large influence on the temperature coefficient. This remains as an important subject for further investigation. The possibility of further reductions in the seismic velocities should also be considered, especially if higher velocities are adopted for the "granitic" layer. Different assumptions as to depth and temperatures of the layers will of course alter these results to some extent.

We have so far said nothing concerning the existence of vitreous phases. The only kind of glass which has been seriously considered and which has also been studied experimentally is glass of a basaltic composition. Adams and Gibson have given the compressibility of one such glass as $14.5 \cdot 10^{-13}$ cm²/dyne, and its initial density, 2.85. Another glass, produced artificially by melting diabase, has been studied in our laboratory[27] and its compressibility determined up to 300° C. In this range, and up to 10,000 kg/cm², the compressibility was about $16 \cdot 10^{-13}$; its initial density was 2.80. The effect of temperature upon the compressibility was small, and indeed at 300° appeared to be abnormal, the compressibility decreasing as the temperature increased. We do not know any other elastic coefficient

[27] Francis Birch and R. R. Law, "Measurement of Compressibility at High Temperatures and High Pressures, *Bull. Geol. Soc. Amer.*, Vol. XLVI (1935), p. 1219.

for this glass. It seems very unlikely that for such material we shall find $\sigma = 0.27$. If we wished to fit it into the intermediate layer, then we might assume that $\sigma = 0.23$, which is about what is given by the ratio of the velocities, 6.3 and 3.7 km/sec. In this case the velocities for Adams' glass are 6.8 and 4.0, and for our glass, 6.5 and 3.8 km/sec. Since these values probably decrease a little as the temperature increases to around 700° or so, they may perhaps be consistent with the velocities in the intermediate layer. Whether this glass would remain stable under such conditions is another question. It appears, however, that the velocities in crystalline and vitreous gabbroic materials tend to become equal as the temperature increases. The absence of new data for the vitreous basalt is due to the difficulty of obtaining samples of sufficient size. Another kind of glass which is sometimes mentioned is vitreous dunite; this glass has never been encountered or produced, except, perhaps, in microscopic amounts.

ACKNOWLEDGMENTS.—Our indebtedness to many collaborators is very great: to Dr. John Ide, who suggested to the authors the possibility of using dynamical methods for studying rocks and whose assistance was invaluable in the realization of this program; to Professor E. S. Larsen for help in the selection of the rocks; to Dr. F. A. Gonyer and Dr. C. S. Hurlbut for analyses; to the Harvard Engineering School for the use of space in the Gordon McKay Laboratory; to Mr. H. J. Ames for the skilful construction of the apparatus; to Professor R. A. Daly for encouragement, criticism, and advice; and to Professor P. W. Bridgman for the development of the high-pressure technique which has made possible this variety of research.

PLUTONIC PERTHITES

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ABSTRACT

Pegmatitic perthites have been studied to a far greater extent than plutonic types. The terminology and theories adequately applied to intergrown feldspars in pegmatites have been indiscriminately transferred to similar kinds in deep-seated igneous rocks with resulting confusion. The present paper is the result of a study of plutonic perthites. Definite relationships exist between bleb shape-sizes and origin. The small ones are due to exsolution; the large, irregular blebs which occur close to and on the margins of the grains, and obviously cut the host, are due to replacement. There are, however, many types that are the result of both processes. Exsolution, at relatively high temperatures, produces small, well-oriented blebs, which at lower temperatures are redistributed, enlarged, and replaced by introduced solutions. Comparison with bleb shape-sizes in pegmatites is made. Complete correlation is not possible at the present time. It may be it does not exist. Criteria for the recognition of various types are given.

INTRODUCTION

There seems to be considerable confusion regarding the origin of perthites in plutonic igneous rocks. Perhaps many petrographers are merely content to recognize that such feldspars are perthitic without attempting to assign different origins to different bleb shape-sizes. However, some do suggest that the perthitic feldspar was, in part, due to exsolution and in part due to replacement, without specifying what "part" was the product of exsolution and what was the result of a replacement process.

Perthites from pegmatites have received a good deal of attention, far more than those in plutonic rocks. The theories of origin of pegmatitic perthites have been applied without sufficient discrimination to plutonic types. My experience has led me to believe that all these theories have a definite place in explaining perthites. Some perthites were formed by one process at a definite temperature range while others were formed by different agents, under different thermal conditions, and still others originated by still other means. I believe that a detailed study of the textural relations and the bleb shape-sizes of plutonic perthites would do much to clear up this confusion. This has been attempted, and this is the justification for this paper.

CLASSIFICATION OF BLEBS

Microscopic examination reveals that there are several common types of perthitic blebs which are distinctive in occurrence, shape, and size to justify the following classification.

Stringlets.—These are the smallest-sized blebs here recognized. They are slender needles with pointed ends. The common size is 0.01 mm. in length, ranging from 0.0075 to 0.0225 mm., with a middle width of 0.00075 mm. ranging from 0.00025 to 0.0012 mm. Under moderately high magnifications (4-mm. objectives) their true shape is not easily available, due to the Becké-line border, because of the difference in the indices of refraction between the stringlet and the host. With reduced light and high powers their shape is revealed. In cross section the stringlets are circular to octagonal in shape. Owing to the double taper, stringlets in thin sections cut at right angles appear to vary in diameter. Stringlets cut near the end are very difficult to see unless especially looked for.

Their distribution is seldom uniform. They are commonly, though not always, confined to the marginal zone of the grains. Other grains show stringlets throughout, but with a greater concentration toward the margins.

Strings.—I formerly thought that the strings were merely large stringlets, but a statistical study has convinced me that strings, though similar in shape, are distinct. They are double-tapering needles commonly 0.06 mm. in length, though they range from 0.02 to 0.09 mm., and in width from 0.0075 to 0.0175 mm., the common width being 0.001 mm. Strings can be compared with stringlets in Figure 1.

The distribution of strings is similar though not identical in behavior to stringlets. They too are usually zoned in respect to the margins, though more grains have strings to the center than is the case with stringlets. The zones of stringlets and strings overlap; they do not intergrade. They are distinct in size and in distribution.

Rods.—These are larger and are the most noticeable type of blebs in plutonic perthites. The basic form is sausage-shaped, round or slightly elliptical in cross section. Variations in shape are common; pointed tips, ribs, "keels," and "fins" usually can be found.

In size they possess a greater range. In length they are commonly

0.06–0.12 mm. with a wide range from 0.02 to 0.19 mm. In width they are commonly 0.02 mm., ranging from 0.01 to 0.05 mm. They

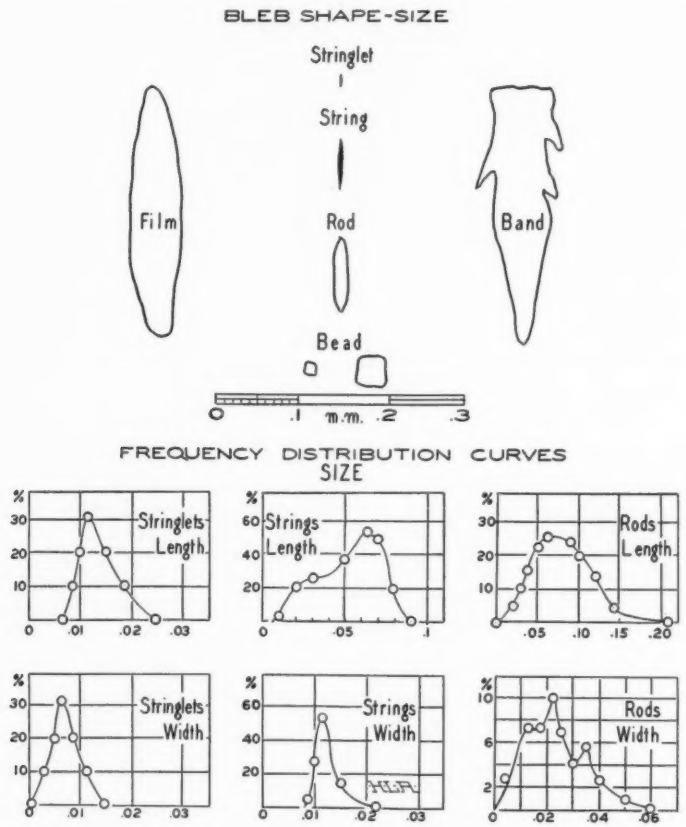


FIG. 1.—Bleb shape-sizes. *Upper portion:* The comparative size and shapes of common plutonic perthitic blebs. All drawn to same scale. *Lower portion:* Frequency-distribution curves of size, for length and for width of stringlets, strings, and rods. The scales are variable, making the curves more uniform. Each based upon many measurements.

are much more irregular in occurrence, in shape, and in size, varying throughout a single grain. Usually they avoid the extreme margin, leaving that free apparently for the strings and stringlets.

Beads.—I am inclined to believe that many of the square-shaped blebs in plutonic perthites are cross sections of rods. Measurements suggest this view. In a few slides, especially oriented and cut for the purpose, this can be proved.

Under high magnifications some, but not all, rods and beads are seen to have been fractured by transverse cracks, resulting in series of plates. Barth[1] has described a similar occurrence in plagioclase blebs in apparently an antiperthite, as follows: "[They consist of] a multitude of very fine disc-like plates or films, one behind the other like the leaves in a book." These cracks are evidently secondary because they cross the rods and pass into the host. A few such fractured rods were seen to have been microfaulted and are illustrated in Figure 2*G*.

Films.—These are, as the name indicates, relatively thin plates, which are irregular in shape and size. In length they vary from 0.1 to 0.5 mm.; in width from 0.01 to 0.15 mm. They behave like veinlets, as some appear to be; they cut the host, usually parallel to the common directions; less frequently they are independent of these, in which case some hydrothermal process is suggested.

Bands.—This term I have taken from Boldyrev[2] and prefer it to Andersen's expression "vein," as it is less likely to cause confusion. These are longer and wider than any of those listed above. They vary considerably from 0.1 to 0.6 mm. or even more in length, from 0.02 to 0.17 mm. in width, and from 0.01 to 0.1 mm. in thickness. Their variations are so great that these figures cannot be as exact as those given for the other types. Band perthites have been observed in plutonic rocks, but they do not show the characteristics suggestive of pegmatites. They are probably due to infiltrations of hydrothermal solutions.

Patch perthite.—When the band types have connecting extensions to produce H-shaped or ladder-shaped joined blebs, the term patch perthite can be used. All gradations between band and patch types exist and at times this classification is difficult to apply. Boldyrev

[1] Tom. F. W. Barth, "Mineralogy of the Adirondack Feldspars," *Amer. Mineral.* Vol. XV (1930), p. 130.

[2] A. K. Boldyrev, "On the Morphology, Genesis and Classification of Perthite and other Intergrown Feldspars," *Trans. Central Geol. and Prospecting Inst. Fascicle 12, 1924*, pp. 1–34. In Russian, with German abstract.

recognizes "sinuous" bands, "rectilinear" bands, "flocky" bands, and "shredded" bands. These adjectives can be applied to the band types as well.

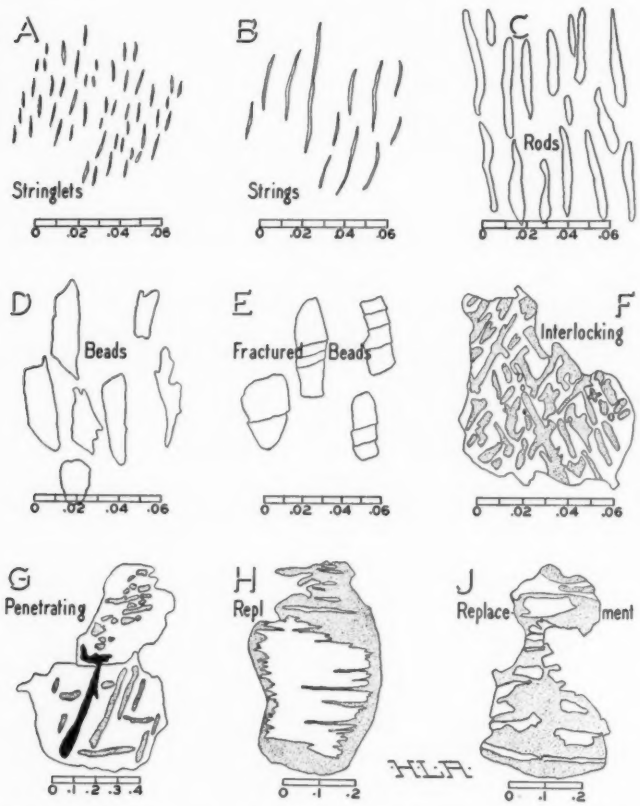


FIG. 2.—Perthitic types. *A*, stringlets from a syenite-granite, Ausable Quadrangle, Adirondacks (1098-Au-22-26); *B*, strings from same; *C*, rods from syenite-granite, Little Falls, N.Y. (1911); *D*, beads from syenite-granite, Little Falls, N.Y. (1905); *E*, fractured beads, syenite-granite, Ausable Quadrangle (1070-Au-15-4333); *F*, penetrating film perthite, syenite-granite, Little Falls, N.Y. (1901); *G*, interpenetrating film, perthite, syenite-granite, Little Falls, N.Y.—the penetrating bleb is emphasized by inking it in black (1901); *H* and *J*, replacing plume, soda orthoclase replacing albite-oligoclase, Ausable Quadrangle (1086-Au-13-244).

Plume type.—There are several additional types listed in my notes which do not lend themselves to clear definitions. However there is a kind that can be called the plume type. When parallel bands or films come together and partially join, they look like feathers. It could be considered a kind of patch perthite.

TEXTURAL RELATIONS

The zonal distribution of the blebs, I believe, is significant in establishing the mode of origin. There are too many grains with the following characters to be mere chance sections. The rod-blebs do not reach the margin of the host-grain. The strings may occupy most of the grain, leaving the margin free. The stringlets are mostly confined to a zone close to the margin, but the extreme margin is usually free. There is no obvious physical connection between the stringlets, strings, and rods with the outside of the grain. The cleavage planes of the host could perhaps have furnished the channelways allowing feldspathic solutions to enter and on crystallization to produce the blebs. But why are there no connecting links? Why is it that the three different types occupy such distinct zones? There seems no reason to believe that the material that forms the blebs came from the outside.

The distribution of stringlet-string-rod blebs is shown in Figures 3 and 4. The measurements for Figure 4 were made with a camera lucida, a cardboard scale beneath the mirror, and a mechanical stage. Each grain was measured from margin to margin. In other words, the diagram is a "cross section," or a kind of "projected profile." Each point is the average of the measurements of ten blebs, of the various kinds, in a line at right angles to the line of section.

There are several significant conclusions to be drawn from these plots. First, the rods are missing in the grain margins. As the center is approached the rods at first are small but acquire greater length. Still farther away from the margin the rods are somewhat smaller. Perhaps this can be expressed better in Figure 4. Second, a grain which is surrounded by different minerals on opposite sides shows an interesting redistribution of blebs. That portion of the perthitic grain nearest an adjoining feldspar encloses more and larger blebs than that portion adjacent to quartz. Thus the kind of adjoining

mineral has a definite influence upon the size and abundance of the rods. In attempting to reach a plausible explanation I have two suggestions: The abundance and larger size of blebs near neighbor-

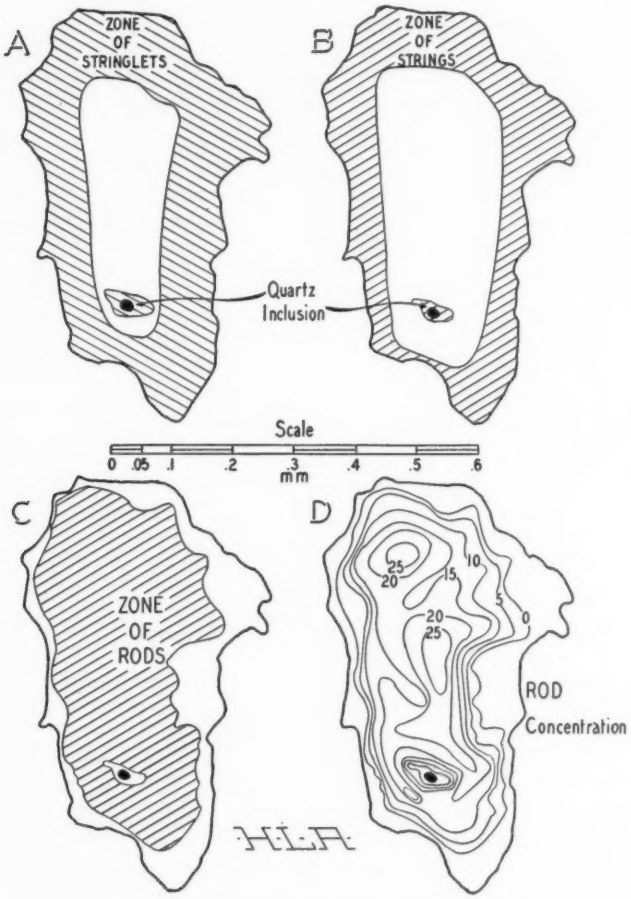


FIG. 3.—*A*, zone of stringlets, nordmarkite, Ausable Quadrangle (1098-Au-22-26); *B*, zone of strings of same grain; *C*, zone of rods in the identical grain; *D*, percentage distribution of rods in the same grain, drawn with a 5 per cent contour interval—note the depression contours around the quartz inclusion.

ing feldspar may mean that that grain of feldspar was the source of the bleb material. Such a view is believed untenable as there are no physical connections from one to the other. There are too many

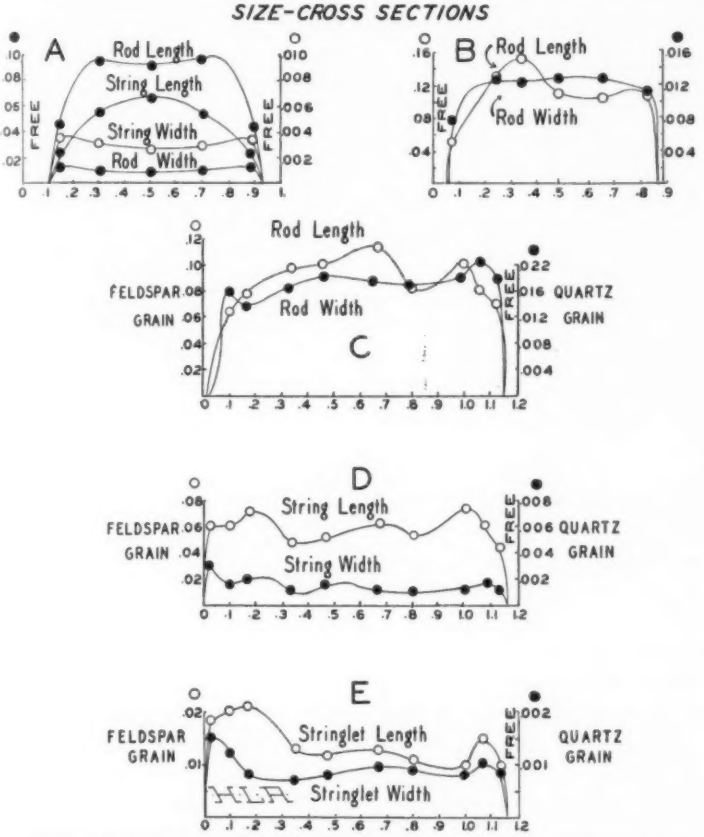


FIG. 4.—Cross-section curves, showing size of blebs in perthites. Horizontal scale is distance; vertical scales, size. *A*, a grain with two "free" margins, Ausable Quadrangle (1048-Au-22-281); *B*, another grain, with one "obstructed" and one "free" margin (1048); *C*, a third grain, rods (1048), with a feldspar grain on left, with narrow "free" margin—a quartz grain is on the right with wide "free" margin in the perthite; *D*, the same grain, strings; *E*, the same grain, stringlets—note greater length near feldspar grain.

such instances to be random sections, which because of their orientation fail to show any connection. The other alternative, the one I entertain, is that the bleb material came from internal sources and was prevented from developing into relatively bulky rods by the constriction due to the viselike grip of the little-contracting adjoining quartz. The difference in the coefficient of expansion and contraction of feldspar and quartz is therefore believed to be the cause.

The acceptance of this explanation implies that the blebs were formed after the crystallization of the quartz and hence this furnishes some idea of the range of temperature obtaining at that time. This further suggests that that portion of the perthitic grain nearest adjacent feldspar is probably older than that close to the quartz. This reasoning is based upon the assumption that in these plutonic rocks the feldspar crystallized before the quartz. This suggests that by noting the distribution of the blebs the older or "obstructed" margin can be distinguished from the newer or "free" margin. Such a grain with "obstructed" and "free" margins has been statistically investigated. The results are plotted in Figure 4*B–E*.

EFFECTS OF QUARTZ INCLUSIONS

Some plutonic perthites contain small inclusions of quartz. They are particularly noticeable because they are surrounded by a "halo" of barren host. Rods apparently never occur near the included quartz. In some instances both stringlets and strings are seen in these cleared rings. In other words, inclusions of quartz produce a "free" area. In Figure 3*D* the results of statistical measurements are given. The diagram is in the form of a contoured surface showing in percentage by area the amount of the rod concentration. It was made by tracing a large (7 × 12 in.) drawing of the grain by a camera lucida. The drawing was then ruled into inch squares and each square measured by a polar-planimeter, area of rods to area of the square. Contours of equal percentages were drawn. Note in Figure 3*D* the necessity of using "depression" contours around the quartz inclusion.

PENETRATING BLEBS

There is an especially significant behavior of an occasional rod or film, which is apparently enlarged and penetrates the barren

zone and even the margin of an adjacent mineral grain. Since barren zones are narrower near feldspar, the mineral usually penetrated is consequently feldspar. This adjacent feldspar is frequently a perthite. Figure 2*G* illustrates an interesting penetrating bleb which has joined one in the neighboring grain producing a T-shaped bleb.

ORIENTATION OF BLEBS

Stringlets and strings are much more uniform in orientation than the rods. The rods not only vary more in shape and size, but also in bending as much as 10° from the directions taken by the strings. Nevertheless they maintain an optical orientation in spite of their physical departures. It is not always possible to secure a crystallographic line of reference within a rod-bleb. Such a direction has to be taken from a twinning plane within the plagioclase rod, if twinning is visible, which is rare, or from the directions taken by the strings alongside the rod. This fact limits measurements to string-bearing perthites in many cases. It is obvious that enlarged, interpenetrating, or interlocking rods, bands, and films depart from the usual directions.

I believe (in conformity with Andersen) that the strings are due to early exsolution, the rods to later exsolution, while the bands and films are still later. These latter forms of perthitic blebs are due either to exsolution or to replacement or to both. The early blebs are more uniform in orientation. The later-formed perthitic blebs "wander." It does not seem to be reasoning a circle to think that the early formed blebs can be distinguished from later forms by the conformity to a regular orientation, the degree of regularity being related to the age. The behavior of blebs at the margins of grains is a criterion independent of orientation.

Andersen[3] sought the cause of the orientation in the directions of minimum, intermediate, and maximum expansion (and contraction) during heating (and cooling) of feldspars. On cooling, feldspars do not contract equally in all directions. The maximum is from 18° to 20° from the (100) direction. The intermediate value is 6°–8° from (001) perpendicular to (010), and the maximum parallel to

[3] Olaf Andersen, "The Genesis of Some Types of Feldspar from Granite Pegmatites," *Norsk. Geol. Tidsskrift*, B. x. h 1–2 (1928), pp. 116–205.

(010). While some blebs are found in accordance with these directions, enough show divergence to indicate that cleavage, parting, expansion, contraction, and orientation of blebs are not always exactly related.[4] There is probably an additional factor involved in the orientation of blebs. Perhaps it may be sought in the following suggestion from X-ray analysis as presented by Bragg:[5]

> Taylor[6] points out that the axes of potash feldspar and albite are almost identical (12.90 A, 7.15 A, and 12.86 A, 7.17 A). This accounts for the lamellae being parallel to (100), since the b and c axes of the two types of crystal coincide in the (100) plane, whereas the a axes running through the lamellae shorten in the albite regions and lengthen in the potash-feldspar regions.

The writer has pointed out previously[7] that blebs occur (1) parallel to (100), which is *not* the direction of maximum expansion, (2) +6.5 to (010), (3) parallel to (001), (4) parallel to (110), and (5) sometimes parallel to $(13.0.\bar{2})$ and (320). The present investigation has confirmed the majority of these directions. No additional ones were found.

CHEMICAL COMPOSITION

Owing to the relatively small size of the blebs in plutonic perthites, determination by optical means is the only feasible method. Extinction angles were used on orientated sections. First it was necessary to draft the diagrams shown in Figure 5*A–C*, which are based upon diagrams by Winchell[8] and Meen.[9] From these plots the approximate compositions have been determined.

It is found that there is great variability in composition. The host, however, is usually a soda-bearing orthoclase, the AB content varying from 10 per cent up to 60 per cent in anorthoclase with occasionally a microcline, and likewise the blebs vary from near albite through oligoclase and andesine to an infrequent labradorite.

[4] H. L. Alling, "Perthites," *Amer. Mineral.*, Vol. XVII (1932), pp. 52–53.

[5] W. L. Bragg, *Atomic Structure of Minerals* (Ithaca, N.Y., 1937), p. 241.

[6] W. H. Taylor, "The Structure of Sanidine and Other Feldspars," *Zeitsch. f. Krist.* Vol. LXXXV (1933), p. 425.

[7] H. L. Alling, "Perthites," *op. cit.*, p. 52.

[8] A. N. Winchell, "Studies in the Feldspar Group," *Jour. Geol.*, Vol. XXXIII (1925), p. 720, Fig. 3.

[9] V. B. Meen, *University of Toronto Geol. Series 35* (1933), p. 43.

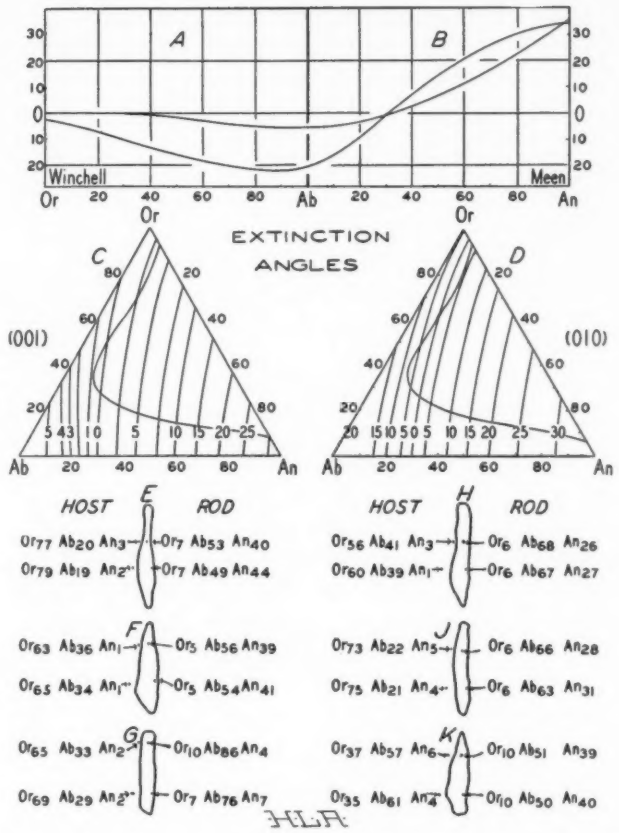


FIG. 5.—*A*, extinction angles of orthoclase-albite after A. N. Winchell, "Studies in the Feldspar Group," *Jour. Geol.*, Vol. XXXIII (1925), p. 720, Fig. 2; *B*, extinction angles of albite-anorthite, after V. B. Meen, *Univ. Toronto Geol. Series 35* (1933), p. 43; *C*, tentative plot of extinction angles, Or-Ab-An system, (100) face—solubility line after E. Mäkinen, "Über die Alkalifeldspäte," *Sonderabd. aus Geol. Fören. Förhandlingar* (1927); *D*, same, for (010) face; *E*, composition of the host close and a little distance from a rod-bleb, and that of center and margin of a rod in perthite in quartz syenite, Saranac Quadrangle (301-Sar-53-425); *F*, same—a different perthitic grain (301); *G*, perthite in granite gneiss, Saranac Quadrangle (330-Sar-54-3729); *H*, perthite in a hornblende syenite, Saranac Quadrangle (396-Sar-42-473); *J*, perthite in a quartz laurvikite, Ausable Quadrangle (1098-Au-22-26); *K*, perthite in a syenite-granite, Ausable Quadrangle (1070-Au-15-4333).

Those rich in albite apparently contain some potash-feldspar. In no case were the blebs or host believed to be simple binary feldspars but were at least ternary in character.

Not only do the blebs and host have variable composition but they vary from point to point within themselves. That is, the center of a bleb has a different composition from the margin. Most, but not all, of the host grains are zoned. So great is the variation that it is difficult to generalize. However, many grains of orthoclasic feldspar have a higher percentage of potash in the center, and richer sodic margins. Likewise, the host varies depending upon the distance from the blebs. Close to plagioclase blebs the host is potash-rich, frequently a potash-albite; farther away the plagioclase contains more lime and becomes an oligoclase. Figure 5*D–J* expresses this more effectively. Consideration of the composition of both blebs and host shows the actual gradational character in spite of the microscopically sharp boundaries between blebs and host. If the blebs had been introduced, would the compositional gradation have occurred? Perhaps diffusion can explain this. If so, it must have occurred at high temperatures, in the magmatic or deuteric ranges, above those of hydrothermal solutions. I do not believe so. The theory of exsolution is perfectly adequate to explain the foregoing facts.

THEORIES OF ORIGIN

Three main theories have been proposed to explain perthitic feldspars: (1) the theory of exsolution, (2) the crystallization of trapped feldspathic liquid, and (3) hydrothermal replacement.

THE PRINCIPLE OF EXSOLUTION

The principle of exsolution is a simple one. Many substances are more soluble at high temperatures than at low. Many rock-forming minerals are solid solutions at elevated temperatures and *super*-saturated at lower ranges. A supersaturated solid solution is not in equilibrium and it separates if it can into additional phases. It is analogous to the separation of crystals of a salt from a hot saturated solution as the temperature falls.

As these minerals are solids and at the same time crystalline, the

expression "crystallize out" is undesirable and, likewise, the word "unmix," which connotes a mixture, is unfortunate. The term "exsolution"[10] is much better, and has been adopted by many geologists.

DIAGRAMMATIC REPRESENTATION OF EXSOLUTION

There are several phase diagrams of the potash-soda feldspars that can be used to explain perthites by exsolution; which one approaches nearest to the truth is impossible to decide at the present time. It may be, as I have maintained, that feldspars of different origins require different diagrams, but this is not the main point. *Exsolution is a natural consequence of any and all of these diagrams.* In Figure 6*G*, which is a typical diagram, feldspar A, at a temperature indicated by T_1, is undersaturated in respect to albite. At the tem-

TABLE 1

COMPOSITION OF FELDSPARS STUDIED BY X-RAYS

Feldspar	Or	Ab	An
1. Adularia, St. Gottard	84.1	8.3	7.6
2. Adularia, St. Gottard	88.3	9.3	2.4
3. Moonstone, Ceylon	74.4	23.1	2.5

perature T_2 the feldspar is saturated and at T_3 it is supersaturated. This feldspar, by inspection of Figure 6*G*, is situated within the field of two solids. To attain equilibrium the supersaturated feldspar (C) must, in time, become a two-phase system, consisting of two solids (C_1 and C_2) each of which is a solid solution. There seems no escape from the implications of the diagram; there must be perthites due to exsolution. The real point is: Can they be recognized from others of different origin?

The X-ray studies of Kôzu and Endô[11] furnish experimental proof of exsolution. Two adularias and one moonstone, with compositions given in Table 1, were X-rayed by the Laue method at various tem-

[10] H. L. Alling, "The Mineralography of the Feldspars," *Jour. Geol.*, Vol. XXIX (1921), p. 222.

[11] Shukusuke Kôzu and Yoshitoshi Endô, "X-Ray Analysis of Adularia and Moonstone and the Influence of Temperature on the Atomic Arrangement of These Minerals," *Sci. Rept. Tohoku Imp. Univ.*, Ser. III, Vol. I (1921), pp. 1–17.

peratures. These specimens were heated to nearly their fusion (softening) points. The two adularias showed a single crystal lattice for the entire temperature range. On the contrary, the moonstone

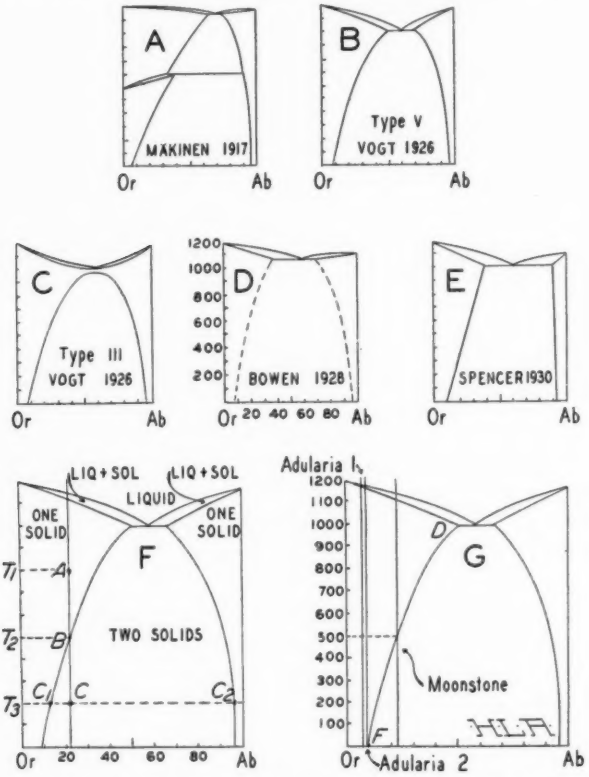


FIG. 6.—*A*, thermal diagram of Or-Ab series, E. Mäkinen, *Geol. Fören. Förhandl.* Vol. XXXIX, No. 2 (1917), pp. 121–84; *B* and *C*, J. H. L. Vogt, *Norsk. Videnskaps Akademi, Oslo I, Mat. Naturvid*, No. 4 (1926), pp. 88–91; *D*, N. L. Bowen, *The Evolution of the Igneous Rocks* (1928), p. 228 ("As the relations are ordinarily to be"). I have added the solubility curves; *E*, Edmondson Spencer, *Min. Mag.*, Vol. XXII (1930), p. 327; *F*, Diagram B, showing details of the process of exsolution—explained in text; *G*, Diagram B with composition of the adularias and the moonstone investigated by S. Kôzu and S. Saiki (*Sci. Rept. Tohoku Imp. Univ.*, Ser. III, Vol. II [1925], pp. 203–38) shown—the line showing the composition of the moonstone crosses the solubility line at 500° C.

showed two crystal lattices, at low temperatures, which became closer together up to about 500° C. when the two lattices became identical and a single set of Laue spots persisted. This shows that the moonstone is a two-phase system at low temperatures and a single-phase one at elevated temperatures. The difference in behavior of the adularias on the one hand and the moonstone on the other is this: They differ in composition and hence occupy a different position on the diagram (see Fig. 6*G*). The two adularias are to the right of solubility line *DF* while the line indicating the composition of the moonstone crosses it at 500° C., below which it is a cryptoperthite and above which it is a homogeneous feldspar.

Still another experimental proof of exsolution is that soda feldspar can be made to dissolve in potash feldspar by heating to temperatures just below melting, if prolonged a sufficient length of time.[12] This is the reverse of exsolution—a kind of re-exsolution.

So far as I know, it was Vogt's[13] paper of 1905 that first emphasized exsolution, although the term itself was not proposed until 1921.[14]

DIFFUSION AND EXSOLUTION

To a physical chemist "diffusion" implies gases and liquids and not necessarily solids. Solids seem so "solid." Nevertheless the phenomenon of diffusion is a real process in the case of metals. Disks of gold and silver, with polished faces, stacked alternately and pressed together for years, have been found to have "grown" together. If sectioned, polished, and examined under a reflecting microscope it is observed that some gold has migrated into the silver and some silver into the gold. The rate of such transboundary migrations is slow, but the rate and the degree of penetration of one metal into another by the process of diffusion increases with temperature and length of time. In Figure 7 this can be seen in the case of copper and nickel. To the metallographer the phenomenon is

[12] E. Dittler and A. Köhler, "Zur Frage der Entmischbarkeit der Kali-Natron. feldspäte, etc.," *Tscherm. Min. Petr. Mitt.*, Vol. XXXVIII (1925), pp. 229–61.

[13] J. H. L. Vogt, "Physikalisch-Chemische Gesetze der Krystallisationsfolge in Eruptivegesteinen," *Tscherm. Min. Petr. Mitt.*, Vol. XXIV (1905), p. 490.

[14] H. L. Alling, "The Mineralography of the Feldspars," *Jour. Geol.*, Vol. XXIX (1921), p. 222.

usually of minor consequence, but to the geologist, with plenty of time to consider, diffusion in solids may be an important process.

Several theories have been proposed to explain diffusion in solids. One is based upon the interchange of atoms from one crystal lattice to the other.[15] This requires that the atoms should "dissociate" from their respective lattices and move across the interfacial boundary. The atoms must become detached from their lattice positions through thermal vibrations. Another theory[16] is that the detached

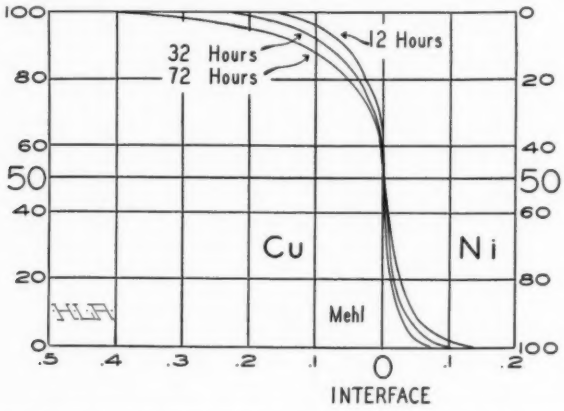


FIG. 7.—Curves showing penetrations of copper into nickel and nickel into copper at 1,000° C. at 12, 33, and 72 hours. Redrawn from Fig. 2 in Robert F. Mehl, "Diffusion in Solid Metals," *Trans. Inst. Min. & Met. Eng.*, Vol. CXXII (1936), p. 16.

atoms can only occupy interstices or imperfections in lattice mosaics. This idea does not seem to have so much support as the first. Langmuir[17] stresses simply *cyclic* interchange of atoms on lattice points. These atoms must possess a minimum energy before they can overcome the potential barrier. This is based upon electrical conductivi-

[15] J. Frenkel, "Über die Wärmebewegung in festen und flüssigen Körpern," *Zeitsch. Physik.*, Vol. XXXV (1926), p. 652.

[16] A. Smekal, *Handbuch der Physik. II*, Vol. XXIV (1933).

[17] Z. Jeffries, "The Trend in the Science of Metals," *Trans. Amer. Inst. Min. & Met. Eng.*, Vol. LXX (1924), p. 303; W. Rosenhain, *Jour. Inst. Metals*, Vol. XXX (1923), p. 3, and "Solid Solutions," *Trans. Amer. Inst. Min. & Met. Eng.*, Vol. LXIX (1933), p. 1003.

ty of salts. Since atoms are completely loosened in molten salts there should be but little difference in their electrical conductivity. Such proves to be the case. The degree of loosening in the solid stage varies greatly, and so does the conductivity.

It has been observed in binary systems of metals that the diffusion coefficients in the end members are greater in the solid solution with the lower melting-point. Also that in binary solids the solubility is greater in the metal with the higher melting-point. Diffusion rates in such systems increase as the melting-points become more divergent and as the solubility in the solid state becomes less. The size of the atom of the diffusing metal and the degrees of symmetry of the metals, involving as they do the factor of anistropy, are variables which are not as yet completely established.

Applying these ideas to exsolution it follows that a supersaturated feldspar, such as *C* in Figure 6*F*, becomes an intergrowth through diffusion. The second phase separates and migrates and forms blebs. Bragg[18] says:

> X-ray analysis of potash feldspar and of albite strongly supports the following explanation of the perthite and cryptoperthite lamellar intergrowth. The framework of linked Si and Al tetrahedra, being similar for the monoclinic and triclinic forms, is continuous throughout the whole structure. At high temperatures the K and Na atoms are evenly diffused in the framework producing a uniform crystal. At lower temperatures there is a segregation by diffusion into potash-rich and soda-rich lamellae producing alternate sheets with monoclinic or pseudo-monoclinic and triclinic symmetry respectively.

CRYSTALLIZATION OF TRAPPED FELDSPATHIC SOLUTIONS

Perhaps another way of expressing this is by the term "simultaneous crystallization." This implies that as the potash-rich feldspar was crystallizing, liquid plagioclase was available and entrapped in the planes of cleavage, but the orientation of the blebs is parallel to (100), which is not a cleavage direction. It may be questioned whether the strength of a growing crystal would have been sufficient to direct and force the crystallizing plagioclase to assume the orientation and shapes observed in the blebs. The chemical gradational nature of the blebs and host is much more suggestive of exsolution than it is of simultaneous crystallization.

[18] *Loc. cit.*

REPLACEMENT

The process of replacement is generally recognized as an important one in the origin of many mineral deposits. It explains a good many things. Solutions react with materials, dissolving some or all of them and depositing others in their places. This means all sorts of changes, ranging from simple dissolving to chemical reaction which produces a more soluble compound that is carried away, making room for the precipitating minerals. This problem raises the question whether it is an atom for atom replacement or whether it is volume for volume. Today it is customary to regard many of these hydrothermal solutions as colloidal and having a thermal range from 573° C. down to perhaps 100° C. Many things can happen in this range.

If potash feldspar is replaced by solutions it is proper to inquire into the nature of such solutions. They are usually albitic. Where did these come from? Are they orthomagmatic or hydrothermal? I should judge from what I have seen that the albite which replaces microcline is the type of albite that Winchell has called analbite,[19] while the albite which replaces orthoclase is apparently the normal type.

Many have stated that perthites are due to replacement even though no evidence is offered and the details of the process are not suggested. Replacement is so well known and so generally accepted that it is easily extended to all perthites. On the other hand, the details of exsolution are better known and yet the principle is not so universally acknowledged. It would be equally misleading to suggest that all perthites are due to exsolution. Some appear to be, certainly not all.

If, during the crystallization of igneous rocks, some soda feldspar should follow the orthoclase (and microcline), the plagioclase could attack and partially replace it and thus produce a replacement perthite. If, on the other hand, the orthoclase crystallized later, replacement antiperthites are the result.

TABULATION OF BLEBS: PLUTONIC AND PEGMATITIC

Barth[20] has introduced the term "guttate." He says it is a textural term, different from any listed by Andersen. His definition is that they are "small drop-like spots, which usually are less than 0.02

[19] *Op. cit.*, p. 720.

[20] *Op. cit.*, pp. 130 and 137.

mm. across." From the microphotograph I should judge that this term includes stringlets, strings, and rods. Some of the latter may be beads. "Guttate," as a term, does not seem to be needed.

Table 2 shows that it is difficult to correlate completely the terms applied to plutonc perthites, on the one hand, and pegmatitic perthites, on the other. Further work may bring about greater harmony, though at present this does not seem so certain. The lack of agreement emphasizes some of the fundamental differences. It is perfectly clear that the *bulk* of plutonic perthites are due to exsolution. There are many illustrations, however, of replacement-perthites in plutonic rocks. In many cases both processes unquestionably operated one after the other. Consequently there are perthites of composite origin. Many pegmatitic perthites in a similar fashion are the result of both processes with, however, a greater emphasis on replacement.

Barth[21] does not take kindly to the theory of exsolution, which he calls secondary "unmixing," in explaining plutonic perthites in certain Adirondack rocks for the following reasons, to which I have added comments:

"1. The basicity of the perthitic substance." I fail to grasp the force of this argument. The solubility line established by Mäkinen and shown in Figure 5*C* and *D* permits blebs to be basic. I have seen what I regard as exsolution blebs as basic as $Or_3Ab_{42}An_{55}$, which is a labradorite.

"2. The fact that the interlocking perthite—which has been formed through a simultaneous crystallization of both feldspars (Andersen . . . .)—grades into the other types." Barth has, apparently, applied a pegmatitic term to a plutonic perthite and hence it may not mean a great deal. My experience with the same kinds of rocks as Barth has examined is that interlocking blebs do not grade into other types. Interlocking types are usually bands and films which I believe to be due to replacement and not due to exsolution.

"3: The fact that all the plagioclases in the pegmatitic nests are of the same composition (which indicates that they all have been deposited from the same liquid, at about the same time)." This may

[21] *Ibid.*, p. 137.

TABLE 2

Bleb Shape-Size Classification

Plutonic (Alling)					Pegmatitic (Andersen)				
Term	Length	Width	Origin	Temp.	Term	Length	Width	Origin	Temp.
Stringlet	0.01 0.0075–0.0225	0.00075 0.00025–0.0012	Exsolution	Very high					
String	0.06 0.02 –0.09	0.001 0.00075–0.0175	Exsolution	Relatively high	String	0.1–1	0.001–0.01	Exsolution	Early stage
Rod and bead	0.08 0.02 –0.3	0.02 0.01 –0.05	Exsolution	Moderately high					
Penetrating film	0.3 0.1 –0.5	0.04 0.01 –0.1	Exsolution and hypothermal replacement	High	Interlocking		0.01	Simultaneous crystallization and exsolution	
Film	0.3 0.1 –0.5	0.06 0.01 –0.15	Hypothermal replacement	High intermediate	Film	Several mm.	0.01	Exsolution	Later stage
Band	0.3 0.1 –0.6	0.08 0.02 –0.17	Mesothermal replacement	Intermediate	Vein	Several mm.	0.1	Recrystallization and slight replacement	
Plume	0.2 0.1 –0.3	0.1 0.01 –0.2	Mesothermal replacement	Low intermediate					
Patch	0.5 0.1 –0.8	0.2 0.05 –0.25	Epithermal replacement	Low	Patch			Replacement	
					Platy				

not apply to plutonic perthites. The uniformity of the composition of the blebs does not necessarily imply a liquid, nor is it an argument against exsolution.

"4. The great variability in the proportions of the individual phases of the perthites." The great temperature range through which perthites are formed presupposes a variation in the degree of separation due to exsolution through the process of diffusion in the solid. In paragraph 3 Barth writes of "uniformity" and in paragraph 4 "great variability." Both can be true, but these facts do not exclude exsolution.

CRITERIA FOR THE RECOGNITION OF VARIOUS KINDS OF PLUTONIC PERTHITES

Exsolution.—Those blebs with comparatively uniform shape, relatively small in size, stringlets, strings, and rods, with a definite orientation, usually parallel to (100) or nearly so; + 6.5 to − 6.5° to (010); parallel to (001); or parallel to (110), which have no connection with the margin of the grain of the host and which avoid zones near inclusions of quartz, are regarded as due to exsolution. The distribution of the rods is a particularly useful criterion. They avoid the margins and zones around quartz inclusions. Exsolution blebs are more abundant and larger near neighboring grains of feldspar and farther away from grain boundaries in contact with quartz, and at the same time they are smaller in size. The stringlets and strings are less restricted in distribution than the rods.

Replacement.—These perthitic guests are larger (films, bands, plumes, and patch), more irregular in shape, do not maintain the same degree of uniformity in orientation, do more obvious cutting of the host, and frequently occur on the margins, sending tongues into the host. In contrast to the behavior of exsolution blebs around inclusions, replacement perthite films, bands, and plumes surround such inclusions.

Sometimes when replacing solutions exercise an attack on a host the contact may be either concave or convex. This is especially true in the case of sulphide ores. These are *caries*, as Lindgren calls them. But their absence does not imply the absence of replacement This

suggestion[22] does not seem to apply to plutonic perthites; evidently the orientation-control of the host is too strong to allow this. I have seen albite replacing microcline with *caries* but that does not produce perthitic textures.

There are many cases of composite perthite, consisting of stringlets, strings, and rods due to exsolution and films, bands, and plumes due to replacement.

CONCLUSION

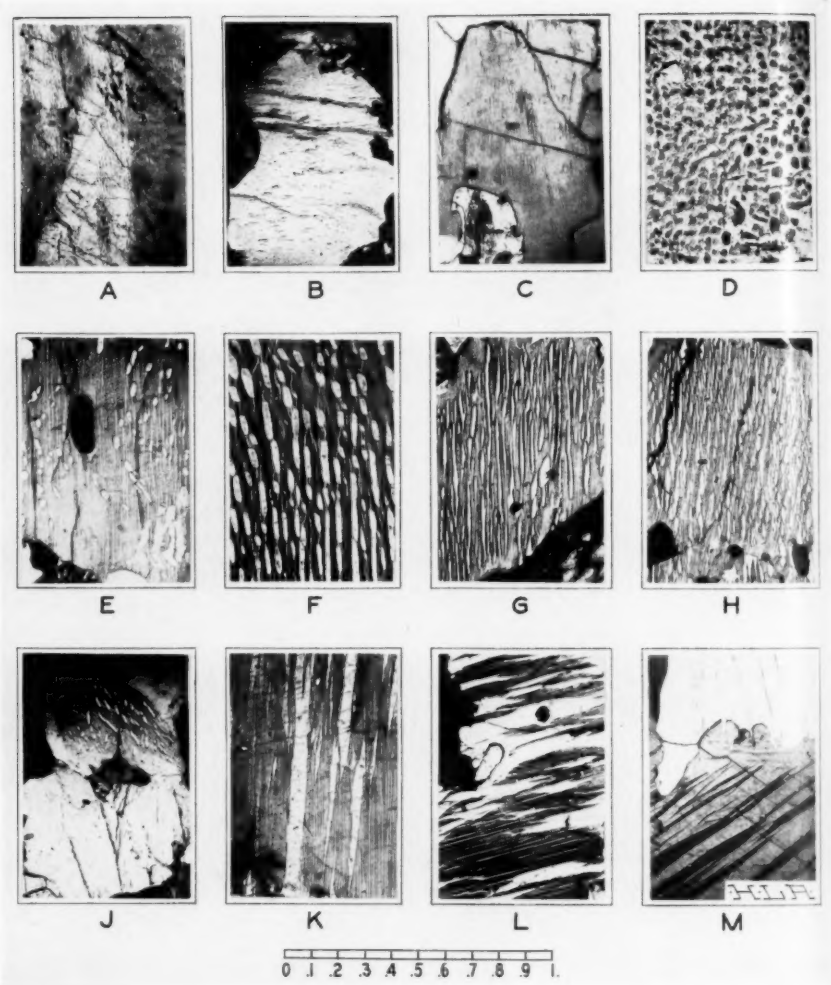
This study has been confined to plutonic perthites; it does not include antiperthites and only refers to pegmatitic types for the sake of comparison.

There is a definite relationship between bleb shape-size and origin. The small blebs, stringlets, strings, and rods are the result of exsolution, accomplished through diffusion in the solid. The larger, irregular blebs which are much less uniform in orientation are mainly due to replacement processes. The element of uncertainty just expressed is actually only apparent because exsolution must have taken place at the start and much of the later replacement was in reality a redistribution and enlargement of the exsolution blebs. There are types of plutonic perthites, formed at comparatively low temperatures, which are largely, and in some cases exclusively, the result of replacement. The fundamental thought is that exsolution and replacement overlap. Simplicity would require the idea of "either or." The facts, however, are against this.

The two extremes—exsolution, a relatively high temperature phenomenon, on the one hand, and replacement, functioning at lower temperatures, on the other—are easily distinguished and recognizable. It is the intermediate types that still need study. The problem is the determination of the amount of emphasis to be placed upon *both* exsolution and replacement. This can be accomplished only through further investigations.

[22] E. S. Bastin, L. C. Graton, W. Lindgren, W. H. Newhouse, G. M. Swartz, and M. N. Short, "Criteria of Age Relations of Minerals," *Econ. Geol.*, Vol. XXVI (1931), pp. 593–94.

PLATE I



EXPLANATION OF PLATE I

(All taken under polarized light, orig. mag. ×90)

A, stringlet perthite, anorthoclase in laurvikite, Laurvik, Norway (609–984)

B, stringlet and string perthite, in quartz laurvikite, Ausable Quadrangle, Adirondack Mountains (611–1098)

C, string perthite, in granite, Ausable Quadrangle (625–134)

D, bead perthite in augite syenite, Long Lake Quadrangle (622–4540)

E, stringlet, string, and rod perthite in granite, Saranac Quadrangle (637–401)

F, stringlet, string, and rod perthite in granite, Saranac Quadrangle (632–330)

G, string and rod perthite in granite, Ausable Quadrangle (621–1048)

H, stringlet, string and rod perthite in granite, Ausable Quadrangle (620–1048)

J, stringlet, string, and rod perthite in quartz laurvikite, Ausable Quadrangle (610–1098)

K, plume perthite in granite; albite-oligoclase replaced by soda orthoclase (which is in part microcline) (618–1086*a*)

L, plume perthite in granite; albite-oligoclase replaced by soda-orthoclase (619–1086*a*)

M, plume—interpenetrating perthite in granite (627–61″).

ON SHALLOW-SEATED FOLDING IN MASSIVE ROCKS[1]

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ABSTRACT

A group of open folds in the San Juan Mountains in southwestern Colorado is described. These involve a basement of pre-Cambrian sediments and granite which was probably buried to a depth not greater than 2,000 feet at the time of the folding (late Triassic or early Jurassic). The presence of a system of joints genetically related to the folds is pointed out, and it is suggested that the folding involved vertical displacements distributed among these fractures. It is also suggested that these displacements were absorbed by plastic flow at greater depths, the transition from flow to fracture moving progressively downward as the deformation increased. In conclusion, an attempt is made to relate the configuration of the folds to broader regional structures.

INTRODUCTION

It is well known that in the folding of stratified rocks a fairly high degree of distortion may be produced by sliding along the surfaces of stratification, accompanied by flexure of the individual beds, without a significant amount of deformation by flow. It is therefore possible for folded structures of this kind to be formed at relatively shallow depths where the confining pressure is not great enough to permit appreciable flow without rupture. In the case of unstratified rocks without pre-existing surfaces of weakness this mechanism is not possible, and the deformation must involve either flow or relative displacement along fracture surfaces. The present paper is a discussion of an example of the latter type which occurs in the southern part of the San Juan Mountains in Colorado. These broad, open folds, of late Triassic or early Jurassic age, were first described by Cross and Larsen[2] as a result of a reconnaissance survey of part of the Pagosa Springs quadrangle. They are exposed in the canyon of the Piedra River where it cuts into the Algonkian basement rocks on the crests of the anticlines.

[1] This paper is developed from a part of the author's thesis for the doctorate in the Division of Geological Sciences, Harvard University (1936).

[2] C. W. Cross and E. S. Larsen, "Contribution to the Stratigraphy of Southwestern Colorado," *U.S. Geol. Surv. Prof. Paper 90E* (1914).

The basement is composed of the schists and quartzites of the Uncompahgre formation and the Eolus granite which intrudes them. Previous to the intrusion the Algonkian sediments were subjected to considerable orogenic disturbance, and the strike of the folds produced at that time is transverse at a large angle to the strike of the Jura-Triassic folds with which we are concerned. Because of their transverse strike and steep dips, the stratification of these sediments had little effect on their mode of yielding to the stresses which produced the later folds.

The thickness of the various pre-Jurassic formations which overlie the basement in this area is approximately as given in Table 1.

TABLE 1

STRATIGRAPHIC SECTION

Age	Formation	Average Thickness (Feet)
Upper Triassic	Dolores	750
Permian	Cutler	400
Pennsylvanian	Hermosa	700
Pennsylvanian	Molas	55
(Upper Devonian or Mississippian)	(Ouray limestone or Leadville limestone)	35
Upper Devonian	Elbert	40
Total		1,980

The figure given for the Dolores is not its present thickness. Much of it was removed during the interval of erosion which followed the folding and preceded the deposition of the Upper Jurassic Entrada sandstone. On the western flank of the San Juan Mountains no structural unconformity separates the Dolores and Entrada, and the average thickness of the Dolores is around 500 feet. This is not necessarily the original thickness, for it is probable that even there some erosion intervened, but west of the mountains, in Paradox Valley, where the amount of erosion was presumably less, and where the initial thickness may have been greater, a thickness of 850 feet has been measured.[3] Granting a considerable factor of un-

[3] R. C. Coffin, "Radium, Uranium, and Vanadium Deposits of Southwestern Colorado," *Colo. Geol. Surv. Bull. 16* (1921).

certainty in this estimate of the original thickness of the Triassic beds, it is highly improbable that the basement was buried to a depth great enough to enable its granite and quartzite to flow to the extent that would be required to produce these folds.

STRUCTURE

The general character of the folding is shown on the map and section (Fig. 1*a* and 1*b*). It will be noted that the formations younger than the post-Dolores folding are also warped to some extent, this warping taking the form of a subdued reflection of the earlier folds. This minor deformation was incidental to the movements which elevated the San Juan dome at the close of the Cretaceous and is of interest in the present discussion in that, when taken in conjunction with the strikes and dips of the Paleozoic beds, it offers additional evidence as to the general trend of the earlier folds beyond their limited areas of exposure. Thus on the northeastern flank of the anticline below Sand Creek the Ouray limestone strikes approximately N. 25° W. and has a maximum dip of 25°–30° NE., and on the southwestern flank it strikes N. 55° W., and dips 10°–15° SW., suggesting that the anticlinal axis has a roughly northwest-southeast trend and plunges toward the southeast. Correspondingly, the average strike of a line along the crest of the post-Cretaceous anticline, as indicated by the structure contours on the base of the Entrada sandstone, is approximately N. 35° W. A similar relationship is found to hold for the other two anticlines. We may, then, use the trends of the younger folds as a rough indication of the strikes of the buried portions of the Jura-Triassic folds with some confidence. On this basis the axis of the truncated anticline at Weminuche Creek strikes N. 32° W., that of the faulted anticline below Sand Creek strikes N. 35° W., and that of the anticline below First Fork Creek strikes N. 75° W.

FRACTURE PATTERN

The basement rocks are cut by closely spaced, smooth, shear joints. In places these divide the rock into slabs an inch or so in thickness, and rarely are adjacent fractures more than a foot or two apart. In order to determine to what extent these joints fall into definite systems and whether these systems have any relation-

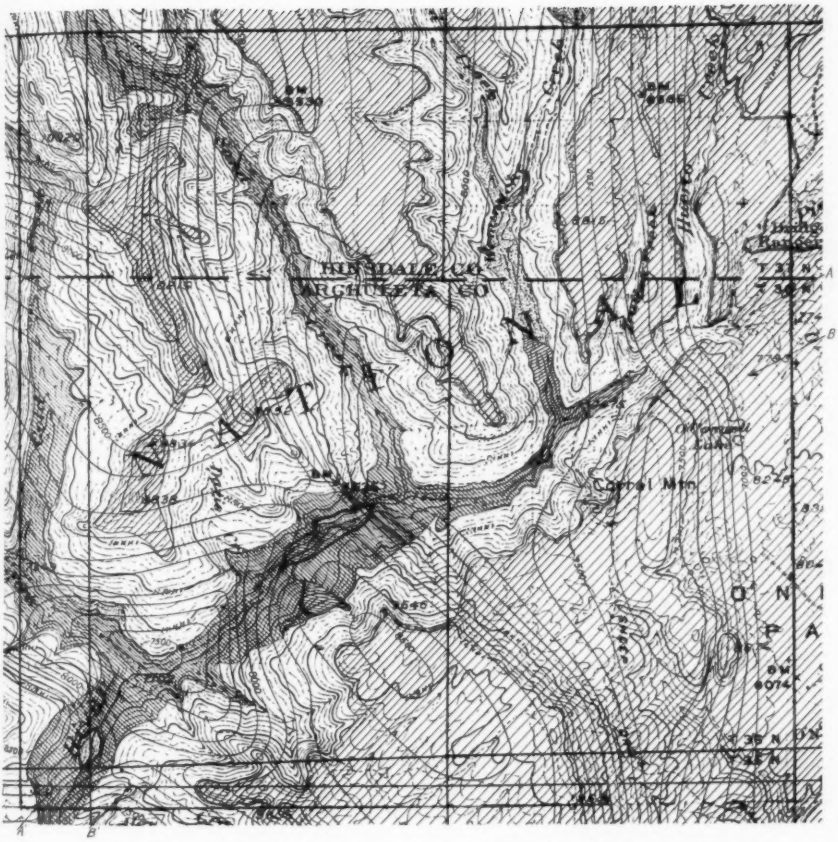

FIG. 1a.—Geological map of part of the Pagosa Springs quadrangle. Structure contours drawn on the base of the Entrada sandstone.

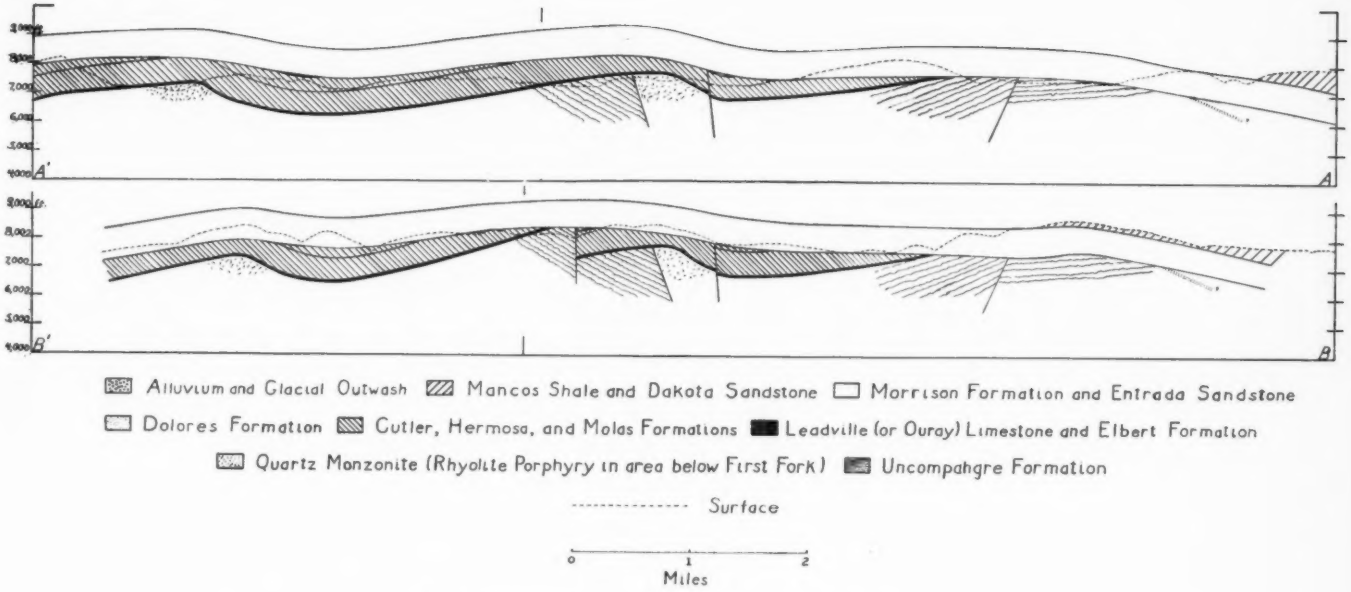

FIG. 1*b*.—Structure sections along lines *A*–*A'* and *B*–*B'*, Fig. 1*a*

ship to the axes of the Jura-Triassic folding, all field observations on the orientation of the joints were plotted on an equal-area spherical projection, each of the three exposed anticlines being treated separately. In each case the existence of two dominant systems of nearly

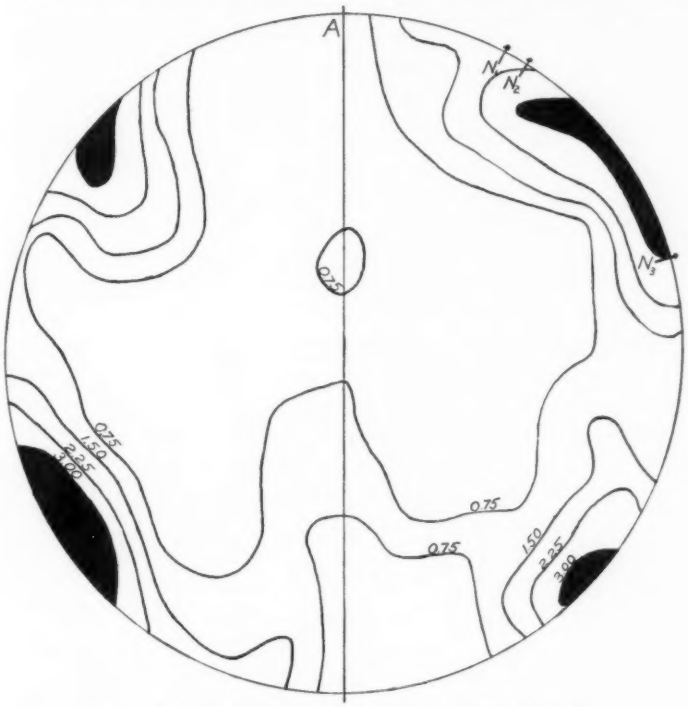

FIG. 2.—Statistical orientation diagram of poles of joint planes

vertical joints became evident through the clustering of the plotted points about the edge of the projection in two different azimuths approximately at right angles to each other. The preferred azimuths are not the same for all three anticlines, which is just what we should expect if the joints are genetically related to the folds, since, as noted above, the fold axes are not parallel but have a divergent arrangement. The preferred orientations are shown by means of statistical distribution contours in Figure 2. This is a composite

diagram constructed by superimposing the three separate point diagrams in such a way as to bring the corresponding areas of maximum point density of each into coincidence and then contouring the resulting composite. On the anticline at Weminuche Creek the line *A* bisecting the angle between the two dominant joint systems strikes N. 25° W., on the one below Sand Creek it strikes N. 32° W., and on that below First Fork it strikes N. 74° W. The agreement between the strikes and those of the anticlinal axes can hardly be a coincidence, and it suggests that, so far as the upper part of the basement was concerned, the folds were produced by relative movement along these two sets of fracture surfaces. There is no relationship between these joints and the strike of the pre-Cambrian folding in the Uncompahgre formation. It is, of course, not contended that fractures related to the older deformation or that primary joints in the granite do not exist, but simply that, of the many joint systems which may be present, two stand out as dominant and that these two are connected with the Jura-Triassic folding. This same mechanism of distributed movement along joint planes has been cited by Wilson in explaining some sharp folding in granite induced by drag on a thrust fault in the Big Horn Mountains of Wyoming.[4]

If the movement along these fractures was distributed more or less uniformly among all members of each system, the amount of relative movement on any one surface needed to produce the observed dips of the flanks of the folds would be of the order of a few inches only. If, instead of being evenly distributed, the movement was concentrated on a few individual surfaces, the amount of relative displacement required would be proportionately greater and, if great enough, should appear in the Paleozoic sediments as small step-faults. Apart from the two major faults below Sand Creek, none was observed. The best exposed horizon in which they should be observable, if present, is the Ouray (Leadville?) limestone. However, the soft shaly Elbert formation, which rarely forms good outcrops, intervenes between the limestone and the basement, and it is quite possible that displacements of the order of a few feet might be so distributed within these incompetent beds as to produce no sharp offsetting in the limestone.

[4] C. W. Wilson, Jr., "A Study of Jointing in the Five Springs Creek Area, East of Kane, Wyoming," *Jour. Geol.*, Vol. XLII (1934), p. 498.

STRESS DISTRIBUTION

The joint pattern that we have described is what one would expect if the applied forces formed a couple acting in a horizontal plane, the axis of the couple being parallel to one system and perpendicular to the other. This situation is represented diagrammatically in Figure 3 for the case in which the axis of the couple is a straight line. Figure 4 indicates what one would expect if the couple acted along a curved line. The similarity of the latter case to the structures under consideration is apparent. It is clear that such a set of forces would not of itself cause the vertical displacements on

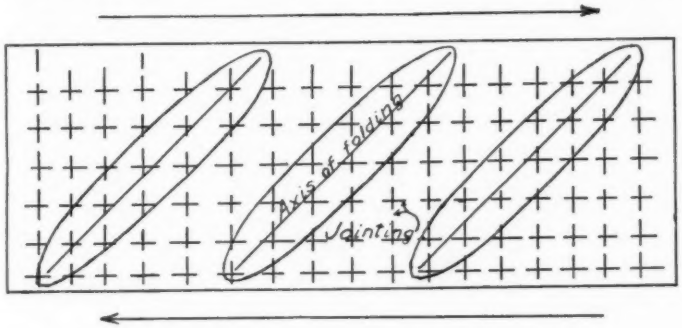

FIG. 3.—Relationship between en echelon folds and joints formed by a rectilinear shearing couple.

these joints which the production of the folds requires. If, however, we suppose that the displacements decreased with depth, the differential movement being absorbed by flow, both the vertical displacements and their apparently even distribution among the fracture surfaces are readily understandable. At the beginning of the folding, when the amount of strain to which the rocks had been subjected was everywhere small, plastic deformation was, presumably, the dominant mode right to the surface. Since, in general, the stress difference required to produce rupture increases with depth, fracturing probably began at the surface and extended downward until it reached a level at which the plasticity was great enough to take up the whole displacement. As the deformation increased, the requisite degree of plasticity would be reached only at correspondingly greater depth. The suggested process might be visualized, then, as the slow

downward propagation of a wave of fracturing, with the upper zone continuously adjusting itself, by slipping along the fracture surfaces, to the form of the folds produced by plastic deformation below the "wave front" of fracture. It should, perhaps, be specifically stated that we are not speaking here of any universally present and sharply separated "zone of fracture" and "zone of flow." If the mechanism we have suggested is correct, any depth at which flow has taken

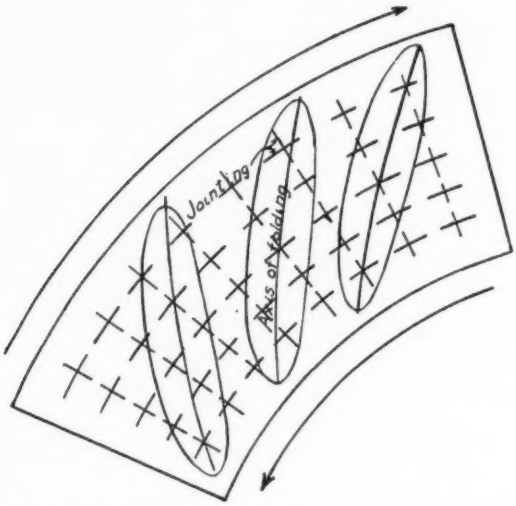

FIG. 4.—Relationship between folds and joints in the case of a shearing couple acting along a curved axis.

place may become the site of fracturing, provided the deformation is carried far enough. We have merely assumed that, other factors being equal, the plasticity of rocks increases with depth, which seems to have ample experimental and observational justification. The term "plasticity" is used here in the sense of the amount of permanent set that can be produced before reaching the rupture point. It is, of course, a function not only of the material, temperature, and pressure but also of such factors as the rate of strain and type of strain.

To complete the discussion, it may be worth while to make a

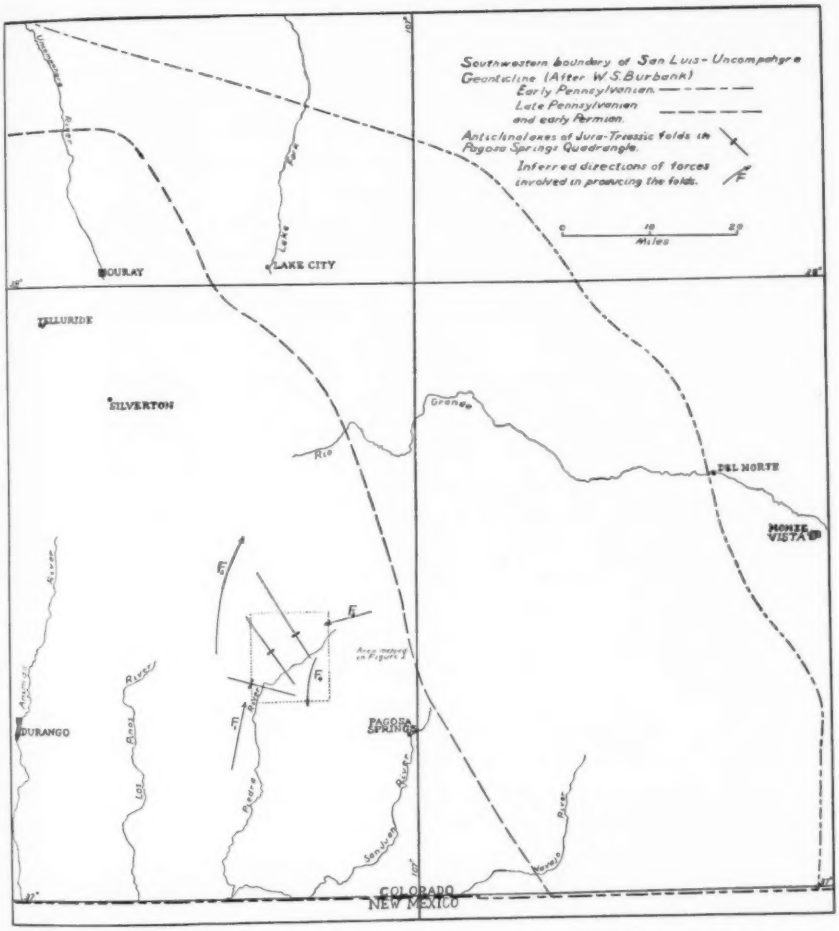

FIG. 5.—Map showing the location of the folds described and the probable distribution of the various forces involved.

frankly speculative and qualitative attempt to relate the local stresses to broader regional causes. In this connection the presence during Permo-Carboniferous time of a mountainous highland to the north and east of the area in which these folds lie may be significant. The existence of this highland has been inferred by Melton[5] from the distribution of coarse clastics in the Pennsylvanian and Permian beds of central and southwestern Colorado and northern New Mexico. It is not unreasonable to suppose that in the late Triassic the basement surface may have stood still higher in the region formerly occupied by this highland than in the area to the southeast and that this may have served to localize the later folding on the flanks of the older structure. According to Burbank's[6] reconstruction of the outlines of this highland, its southwestern boundary formed an arc around the eastern and northern border of the Pagosa Springs quadrangle, the apex of the arc being near the northeastern corner. There might be a tendency for a thrust acting from the south, or a little west of south, to be deflected toward the east into this re-entrant. If the material displaced toward the apex in the center of the arc moved farther than that held back by tangential resistance on the sides, a horizontal shearing couple would be developed. This situation is represented diagrammatically in Figure 5, the arrow F_1 representing the active thrust; F_3 and F_4, the resistance exerted by the eastern side of the arc; and F_2, the forward drag exerted by the material in the center displaced toward the apex. If this is a correct interpretation, the area of maximum deformation must have lain north of the area we have described, but this cannot be verified, since the Paleozoic sediments have been removed from those parts of the San Cristobal and Needle Mountain quadrangles in which the extension of these folds lie.

ACKNOWLEDGMENT.—The writer is indebted to Professor E. S. Larsen both for the original suggestion of the problem and for helpful criticism in the course of its development.

[5] F. A. Melton, "The Ancestral Rocky Mountains of Colorado and New Mexico," *Jour. Geol.*, Vol. XXXIII (1925), p. 84.

[6] W. S. Burbank, "Relationship of Paleozoic and Mesozoic Sedimentation to Cretaceous-Tertiary Igneous Activity and the Development of Tectonic Features in Colorado" in *Ore Deposits of the Western States* (A.I.M.E. Lindgren volume [1933]), p. 277.

THE OSTEOLOGY OF *AULACOCEPHALODON PEAVOTI* BROOM

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AND
FRANK BYRNE
University of Chicago

ABSTRACT

The skull and skeleton of the anomodont reptile *Aulacocephalodon peavoti* are figured and described. A restoration based on the mounted skeleton in Walker Museum is presented.

INTRODUCTION

The purpose of this paper is to figure and describe the skull and skeleton of the specimen of *Aulacocephalodon peavoti* mounted in Walker Museum, University of Chicago. This specimen was originally identified as *Dicynodon* (*Aulacocephalodon*) *tigriceps*, and, under this name, a preliminary account was given by Olson.[1] Upon reexamination it appears that the specimen must be referred to *Aulacocephalodon peavoti* Broom, a closely related species. This species was named by Broom in 1921,[2] and at that time placed in a new genus, *Bainia*. Subsequently Broom, recognizing the priority of Seeley's genus *Aulacocephalodon*, referred his species to that genus.

The specimen described in the present paper, University of Chicago, No. 1532, was found on the Toverwater farm near Murraysburg, South Africa, in the lower *Cistecephalus* zone.

DESCRIPTION

SKULL

A well-preserved skull of this species of *Aulacocephalodon* was described by Broom[3] in 1921 and refigured by him in 1933. His account and figures of the dorsal and lateral surfaces are quite com-

[1] E. C. Olson, "A Mounted Skeleton of the Mammal-Like Reptile *Dicynodon tigriceps*," *Jour. Geol.*, Vol. LXIII (1935), pp. 1063–66.

[2] R. Broom, "On Some New Genera and Species of Anomodont Reptiles from the Karroo Beds of South Africa," *Proc. Zoöl. Soc. London* (1921), pp. 659–61.

[3] *Ibid.*

plete and accurate and nothing can be added to them from the specimen under consideration. The general shape of the new skull and the arrangement of the elements, so far as this can be determined, are shown on Figures 1*A* and 2*A*.

The palate has been exposed but the sutures are so obscure that the limits of most of the elements cannot be determined. The essential features of this portion of the skull are shown in Figure 1*B*. The maxillaries and premaxillaries are toothless, and there is no indication that any teeth were ever present. In the canine region of the maxillary, however, there is a large, rough projection of bone, which, if covered with a tough substance, would have acted as an efficient substitute for teeth.

TABLE 1

PRINCIPAL SKULL MEASUREMENTS

Greatest length.....	470 mm	Interorbital width...	105 mm
Greatest width.....	410	Intertemporal width.	85

On the posterior part of the figure of the palate the positions of the external openings of the fallopian canal, the fenestra ovalis, and the jugular foramen are indicated. The fallopian canal for the seventh cranial nerve is large. It appears to lie just above the suture between the proötic and basisphenoid. The fenestra ovalis, which lies back and slightly medial to the opening for Nerve VII, is rather small for such a large skull. The stapes is absent. The jugular foramen is large and appears to lie between the basioccipital and exoccipital.

The brain case has been partially exposed but the preservation is such that only a general description of it is possible. The cavity of the brain is very high and narrow. The anterior margin of the proötic is nearly vertical and is marked by a very shallow proötic incisure and a poorly defined dorsal, venous notch. Anterior to the proötic, the supraoccipital and the parietal project ventrally to form the lateral walls of a groove leading from the otic portion of the brain case to the pineal opening. It is impossible to determine the structure anterior to the pineal opening.

The basisphenoid is a broad, deep bone marked ventrally by strong tubera. The rostral portion, the parasphenoid, projects for-

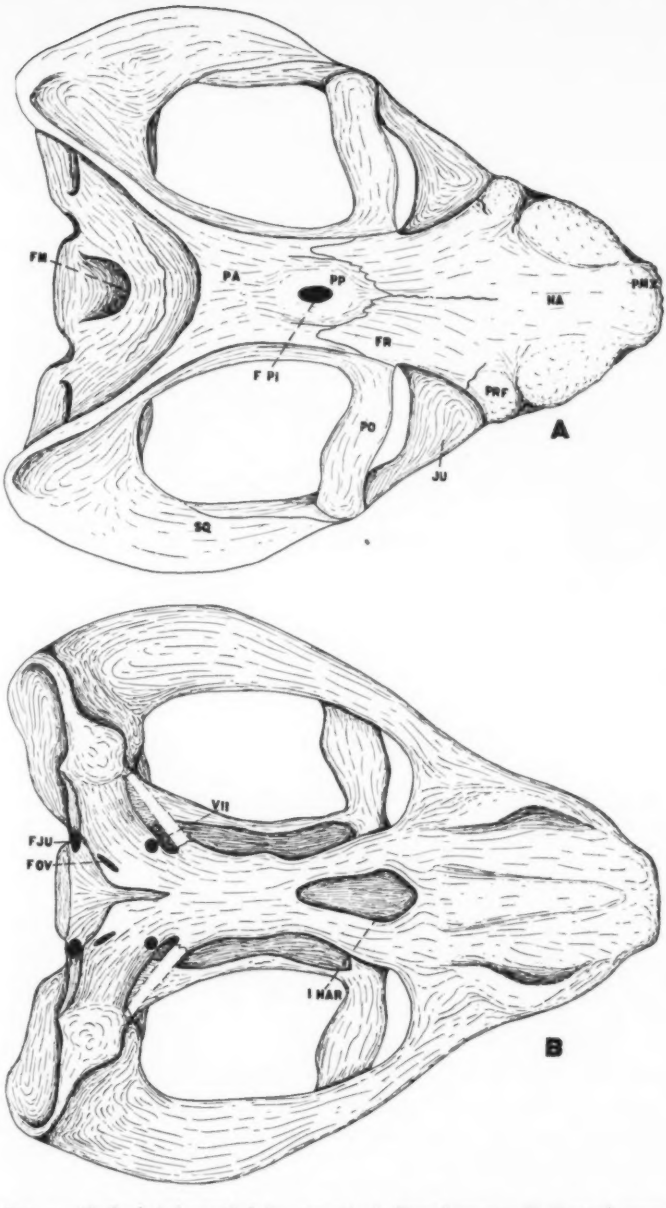

FIG. 1.—Skull of *Aulacocephalodon peavoti*. *A*, Dorsal aspect; *B*, Ventral aspect. *F JU*, jugular foramen; *FM*, foramen magnum; *F OV*, fenestra ovalis; *F PI*, pineal foramen; *FR*, frontal; *IN*, internal nares; *JU*, jugal; *PA*, parietal; *PMX*, premaxillary; *PO*, postorbital; *PP*, preparietal; *PRF*, prefrontal; *NA*, nasal; *SQ*, squamosal; VII, fallopian canal for cranial nerve VII. × about ¼.

ward and slightly upward to about the level of the pineal opening. The pterygoid lies beneath the anterior portions of this bone.

LOWER JAWS

The lower jaws are essentially complete and their shape, shown in Figure 2*B*, is quite clear. The sutures between the elements are very obscure and only a few can be made out with any certainty. The

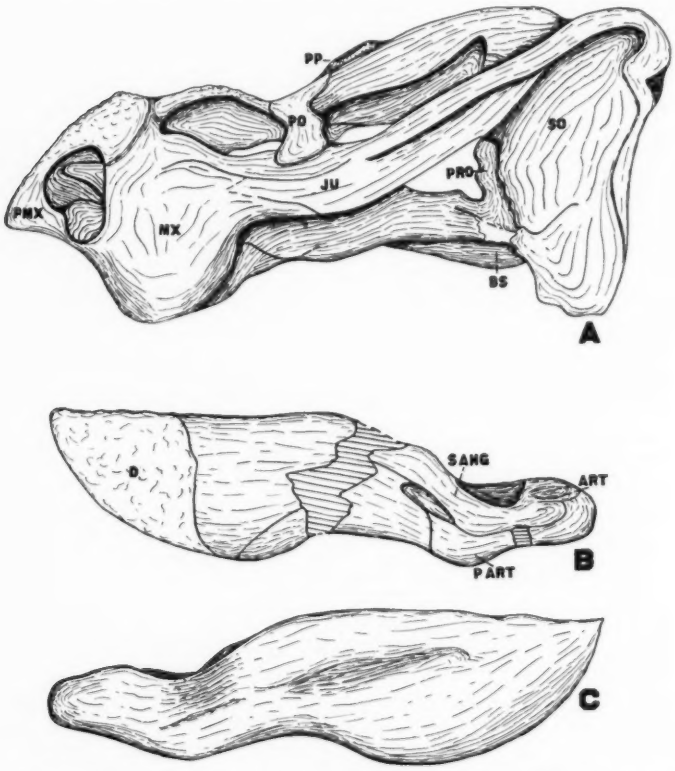

FIG. 2.—Skull and jaws of *Aulacocephalodon peavoti*. *A*, Lateral aspect of skull; *B*, Medial aspect of right lower jaw; *C*, Lateral aspect of right lower jaw. *ART*, articular; *BS*, basisphenoid; *D*, dentary; *JU*, jugal; *MX*, maxillary; *P ART*, prearticular; *PMX*, premaxillary; *PO*, postorbital; *PP*, preparietal; *PRO*, proötic; *SANG*, surangular; *SQ*, squamosal. × about $\frac{1}{4}$.

anterior ends of the lower jaws are very heavy and meet in a strong symphysis. The posterior portions are light and the articulation with the skull is not particularly strong.

VERTEBRAL COLUMN

The twenty-six presacral vertebrae present in the specimen appear to form a complete series. Because they are somewhat distorted it is impossible to determine the exact number of cervical vertebrae, the vertebrarterial canal having been obliterated. Likewise it is impossible to be certain of the exact number of thoracic and lumbar vertebrae.

The atlas-axis complex is well preserved. The atlas consists of a crescent-shaped intercentrum, which is closely applied to the lower half of the odontoid, and of two separate dorsal portions. The dorsal segments are wing-shaped structures, pointed medially but expanding anteriorly and posteriorly so that, viewed in superior aspect, the outline is wedge-shaped, the base of the wedge occupying a lateral position. From the anterior half of the base of the wedge arises the pointed transverse process. This may also include the first rib. The anterior border of the process is curved ventrally to form a well-developed flange.

The centrum of the first vertebra is fused firmly to that of the axis. The odontoid process so formed is moderately convex. The centrum of the axis is elongate and concave laterally. The neural arch rises above it to enclose a restricted neural canal. The neural spine is quite high and is expanded anteroposteriorly. The anterior border of the spine curves markedly forward. The transverse processes, arising from the anterior half of the neural arch, take origin immediately above the centrum and curve sharply ventrad.

Because of the distortion of the vertebrae, the description of the remainder of the presacral region can be only very general. Throughout this region the centra are rather cylindrical and only slightly concave on their sides. There is a uniform increase in size from anterior to posterior. The most massive centrum is that of the last presacral vertebra, this being almost double the size of the centrum immediately following the axis.

The neural spines are all about the same height. There is, how-

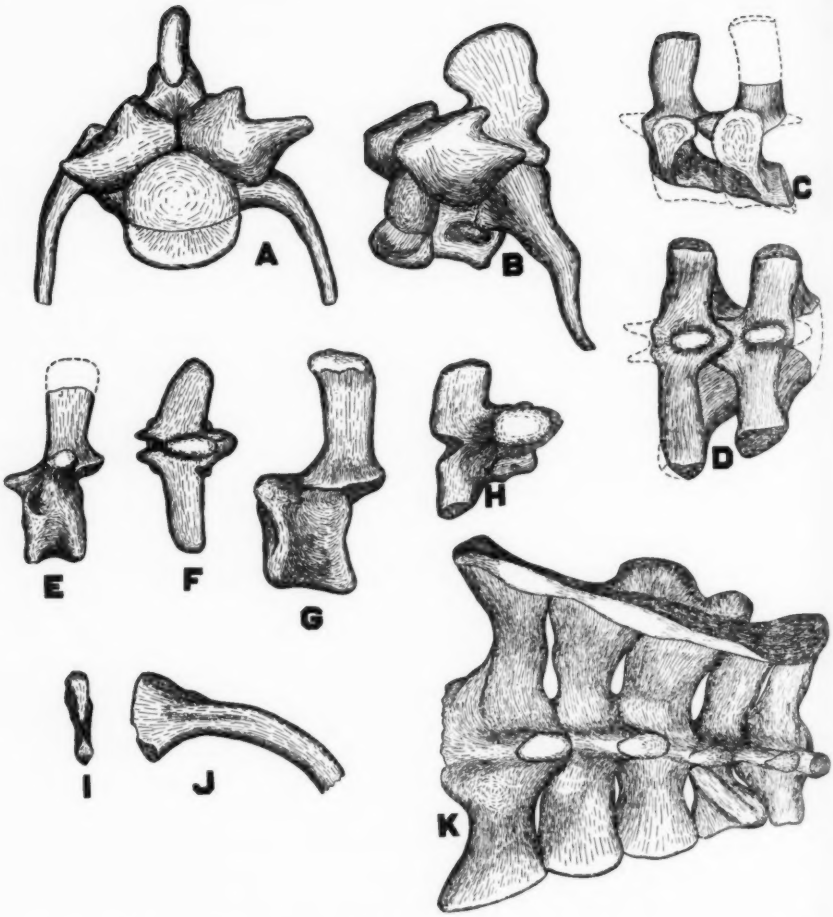

FIG. 3.—Vertebrae and ribs of *Aulacocephalodon peavoti*. *A*, Atlas and axis, anterior aspect. *B*, Atlas and axis, lateral aspect, left. *C*, Vertebrae 6 and 7, lateral aspect, left. *D*, Vertebrae 6 and 7, dorsal aspect. *E*, Vertebra 15, lateral aspect, left. *F*, Vertebra 15, dorsal aspect. *G*, Vertebra 21, lateral aspect. left. *H*, Vertebrae 21, dorsal aspect. *I*, Rib 7?, head in proximal aspect. *J*, Rib 7?, anterior aspect. *K*, Sacrum, dorsal aspect. $\times\frac{1}{4}$.

ever, an increase in massiveness posteriorly. The increase is the result of an anteroposterior expansion of the spines. The spines tend to become bulbous toward their summits. The spines in the cervical region appear to be directed slightly to the posterior, those in the thoracic region to stand more or less vertically, and those in the general lumbar region to be directed slightly to the anterior.

An interesting migration of the transverse processes is to be noted. At the anterior end of the column the transverse processes arise from the base of the neural arch and from about the center of that structure. Posteriorly the processes raise somewhat higher on the arches. At the same time their origins move progressively forward, until, in the lumbar region, the transverse processes spring from the sides of the anterior zygapophyses.

The sacrum includes five vertebrae all closely opposed to one another, although apparently not ankylosed. There is a gradual decrease in size from the first sacral to the fifth. The neural spine of the first sacral curves somewhat anteriorly while those of the remaining four are directed posteriorly. The transverse processes are very short.

Except for five centra the caudal region has been entirely restored. The centra preserved are short anteroposteriorly and roughly circular in transverse section. Apparently the transverse processes arise at the points of junction of the centra and the neural arches.

RIBS

The ribs are dichocephalous throughout the presacral region. Posteriorly, however, the capitulum rises until, in the last presacral, it is very close to the tuberculum. All the tubercular heads attach to the ventral sides of the transverse processes. Anteriorly the capitular facets are centrally located. Posteriorly they are shifted in position so that a single facet becomes two demifacets, the anterior demifacet being situated on the posterior border of the preceding centrum.

The presacral ribs taper gradually toward their ventral extremities. The cross sections are oval proximally and slightly elliptical distally. The arch of the ribs indicates a fairly spacious body cavity. The longest ribs are those of the middle of the section.

The sacral ribs are all firmly fused to the transverse processes. The first sacral is the most massive of the series, the other four becoming successively shorter and less massive. The middle portion of each rib is somewhat constricted, the internal portion flaring broadly to its articulation with the transverse process and the external portion flaring broadly to its articulation with the ilium. The first sacral rib, oval in section, descends sharply to the anterior.

PECTORAL GIRDLE

The preserved portion of the shoulder girdle consists of a nearly complete right scapula, a left scapula from which only the anterodorsal corner is missing, the left post-coracoid, and the anterior portion of the interclavicle. In the accompanying figure and the restoration the precoracoids, the clavicles, the cleithrum, and a sternum have been added.

The scapula is a bladelike structure expanded along its superior border and tapering to its greatest constriction just above the glenoid. The posterior border is markedly concave, the anterior border almost straight. The scapula is arched externally in a dorsoventral plane to accommodate itself to the expansion of the thoracic cavity. The axillary ridge causes the posterior border to be somewhat thickened. The ridge appears slightly below the superior border and proceeds to the glenoid buttress with which it merges. The scapular border of the glenoid is directly somewhat posteriorly. A strong glenoid buttress is present. A well-developed acromion process appears slightly below the center of the scapula on the anterior border.

The postcoracoid exhibits a strong ridge running from the glenoid to the posterior corner. Otherwise the element is thin and platelike. The glenoidal surface is produced posteriorly, is slightly concave, and is directed dorsally and laterally more than it is posteriorly. The anteroventral position of the coracoid, as well as the precoracoid, seems to be missing.

The interclavicle is greatly expanded along its anterior border and sends out winglike flanges for its contacts with the clavicles. Posteriorly the interclavicle constricts sharply to a stout, slightly curved projection, which presumably underlay the sternum.

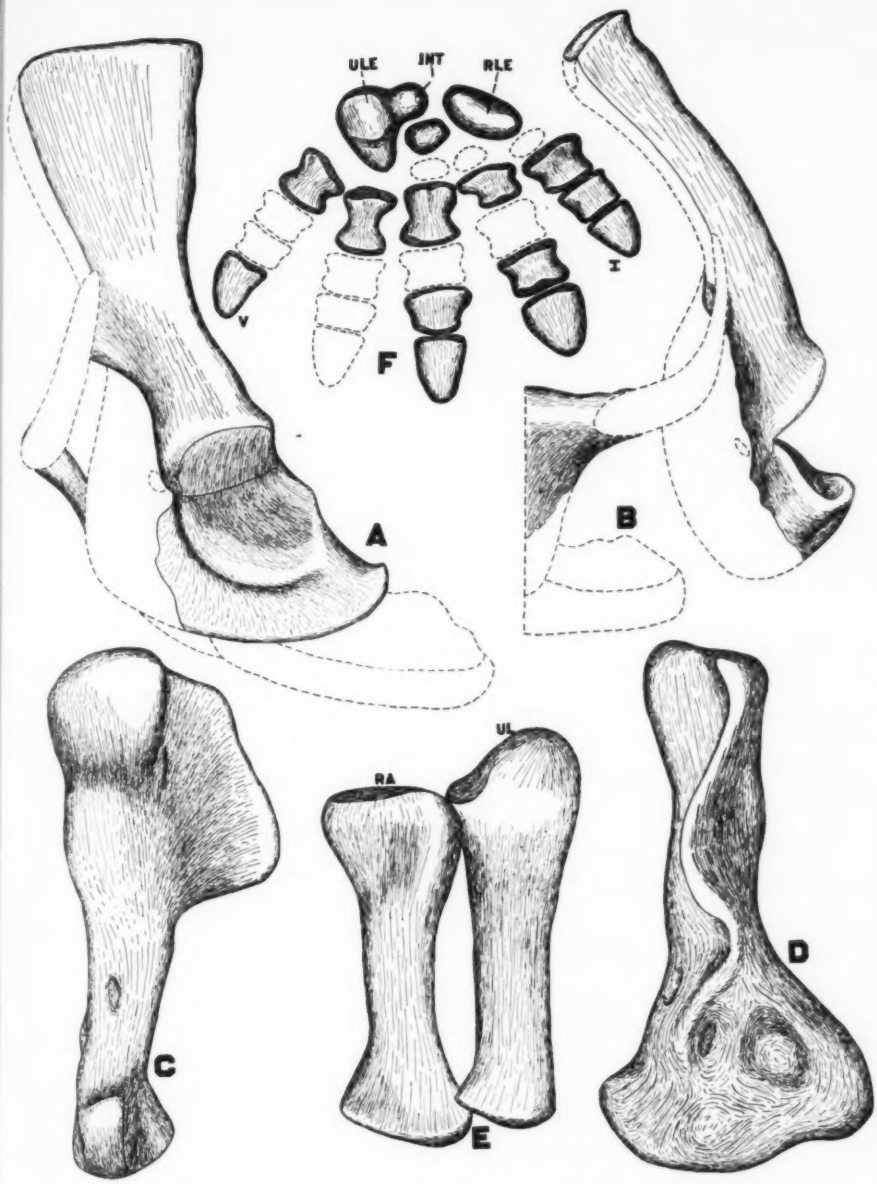

FIG. 4.—Pectoral girdle and forelimb of *Aulacocephalodon peavoti*. *A*, Left side of pectoral girdle, lateral aspect. *B*, Left side of pectoral girdle, anterior aspect. *C*, Left humerus, posterior aspect. *D*, Left humerus, ventral aspect. *E*, Left radius and ulna, anterior aspect. *F*, Forefoot, right. *INT*, intermedium; *RA*, radius. *RLE*, radiale; *UL*, ulna; *ULE*, ulnare. ×½.

FORELIMB

The forelimb as preserved includes both humeri, the radii and ulnae of both sides, and a considerable portion of the right wrist and hand.

The humerus is a very powerful element. A large, nearly round head gives origin from its anterodorsal border to a greatly expanded deltopectoral crest. At the point of subsidence of the crest a short but stout shaft appears. The shaft soon expands distally to form the massive trochlea and somewhat less massive capitulum. The epicondyles are but slightly developed. A conspicuous entepicondylar foramen is present. There is no indication of an ectepicondylar foramen. The plane of the distal condyles lies at an angle of about 30° to that of the medial border.

The radius and ulna are short but massive bones. Proximally each one is subcircular in cross section. The shafts are considerably flattened in the lateral plane. The distal end of the radius flares both medially and laterally while that of the ulna is expanded only medially. The ulna bears a massive, rounded olecranon process.

The pisiform and the distal carpals are missing from the wrist. The radiale is an elongated, egg-shaped element. The ulnare is flattened on its dorsal and ventral surfaces. Viewed from above, the ulnare is nearly circular in outline. Apparently fused to the medial border of the ulnare is the small, spherical intermedium, and fused to its distal border, a small, pointed second centrale. The first centrale is small in size and ovoid in form.

All the metacarpals are present, and all conform to a single general pattern. The proximal and distal borders are expanded and somewhat convex. The waist is conspicuously constricted.

The first phalanx of the first digit is a short element and is somewhat flattened dorsoventrally. The ungual phalanx is short and bluntly pointed. The first phalanges of the second and third digits are missing, the second phalanges quite short, and the ungual phalanges bluntly pointed. The fourth digit is missing and only the medially curved, moderately pointed ungual phalanx is preserved of the fifth digit. The phalangeal formula appears to be 2–3–3–3–3.

PELVIC GIRDLE

The right side of the pelvic girdle has been preserved intact except for the medial portions of the pubis and ischium. The left side of the girdle is absent.

The ilium consists of an anteroposteriorly expanded dorsal flange rising above the stout ventral shaft. The superior border, curved throughout most of its length, passes sharply downward anteriorly and less markedly so posteriorly. The anterior process is longer than the posterior. The anterior half of the external surface of the crest is somewhat concave; the posterior half, slightly convex. Anteriorly the crest flares laterally.

The greatest anteroposterior constriction of the ilium is slightly below the crest. Ventrally from that point the element expands considerably to contribute to a capacious acetabulum. A strong acetabular buttress is developed on the anterior half of the ilium.

The acetabular portion of the pubis is quite thick. Anteriorly it bends sharply toward the median line. Posteriorly it slopes ventrally to its contact with the ischium. Only the external margin of the puboischiadic foramen is preserved.

The ischium, markedly concave along its vertically directed posteroexternal border, curves sharply toward the median line. The element lies principally in a nearly vertical plane.

The dorsal half of the acetabulum is developed from the ilium. About two-thirds of the ventral half are contributed by the ischium. The iliac portion faces both ventrally and externally, the ischial and pubic portion, only externally.

HINDLIMB

The femur, tibia, a portion of the fibula, and the proximal tarsals are preserved on the right side. Of the elements of the left leg, only the fibula is preserved.

The femur is a powerful, slightly flattened element. The head is poorly differentiated from the greatly expanded greater trochanter, which does not subside into the shaft for fully half the length of the bone. The expanded distal end is composed of poorly distinguished condyles.

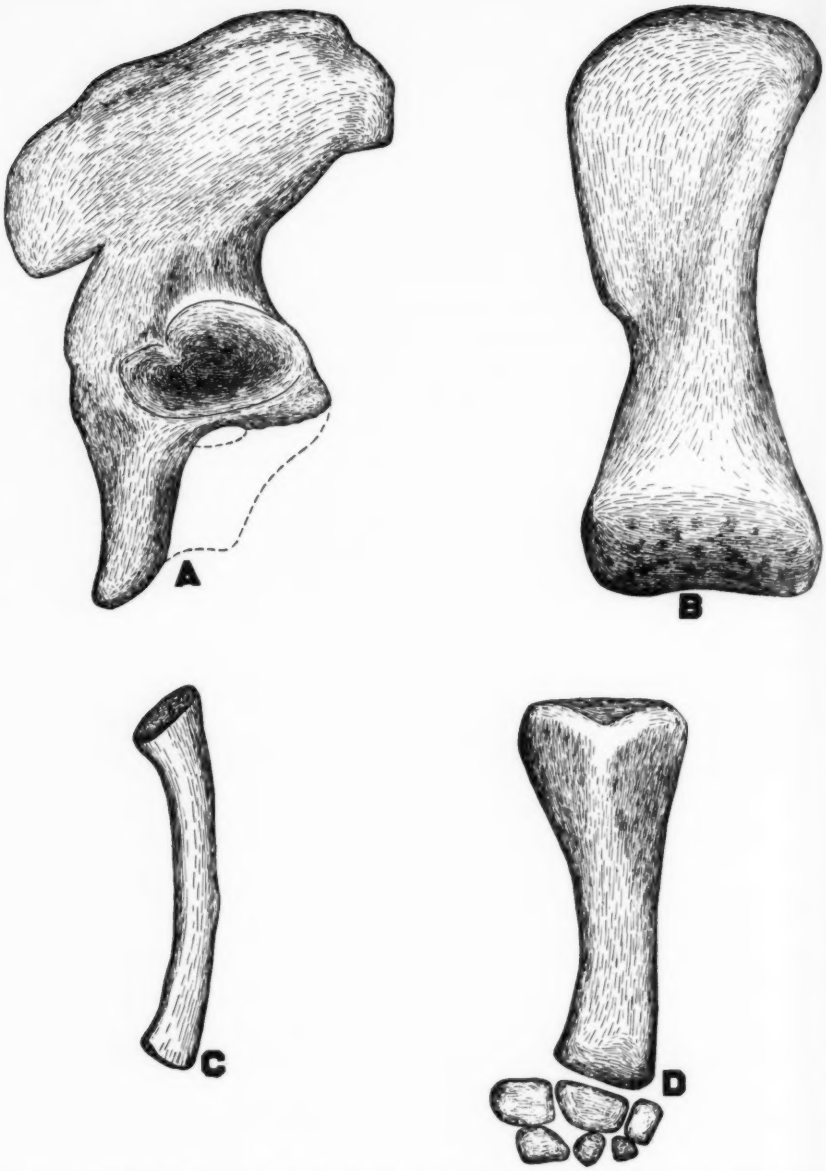

FIG. 5.—Pelvic girdle and hindlimb of *Aulacocephalodon peavoti*. *A*, right side of pelvis, lateral aspect. *B*, right femur, anterior aspect; *C*, Left fibula, anterior aspect; *D*, Right tibia and proximal tarsals, anterior aspect. ×½.

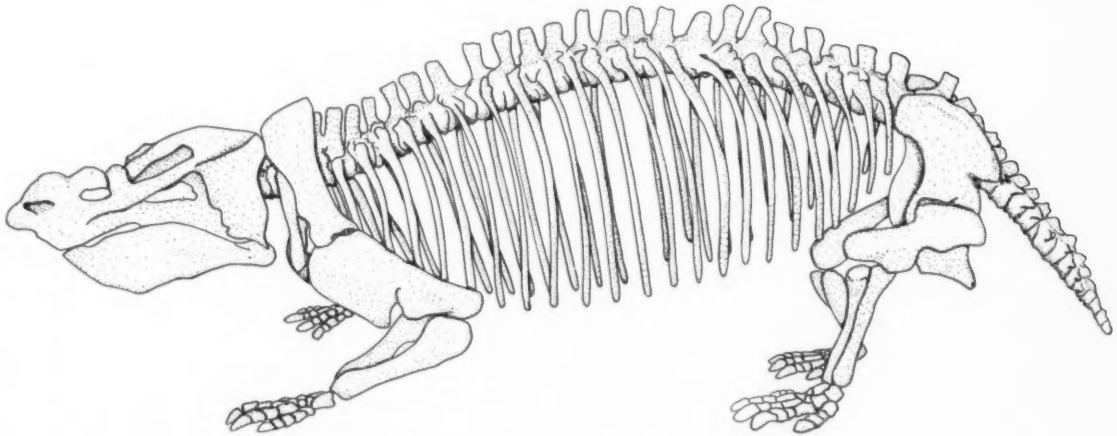

FIG. 6.—Restoration of *Aulacocephalodon peavoti* Broomi

The stout tibia presents an expanded proximal articular surface to the femur. Distally it tapers to a somewhat flattened shaft. The internal border is markedy concave, the external border nearly straight. The fibula is a slender bone roughly circular in cross section. The proximal end is but slightly expanded, the distal end even less so. The element is strongly concave along its internal border.

The ankle and hind foot are missing in the specimen except for some of the proximal tarsals. The elements preserved may be identified as a small, square intermedium; a five-sided, fairly large tibiale; two small centrales; and two other elements which might be the fibulare and a somewhat smaller navicular.

NOTES ON THE RESTORATION

The accompanying restoration of *Aulacocephalodon peavoti* is redrawn from photographs of the mounted specimen and shows the attitude of the skeleton as mounted. Distortion of the bones imposed certain limitations on the positions in which they could be placed in the mounted skeleton and a few portions are not precisely in the positions which they probably assumed during the life of the animal. Most striking in this regard is the shoulder girdle, which probably is too far forward, mainly a result of the forward projection of the scapula. The forelimb perhaps actually was somewhat more under the animal, although the nature of the articulation of the girdle and humerus, and the humerus and radius and ulna, indicates that the limb did project laterally. The ribs were distorted from their original form so that the ventral extremities project more directly downward than probably was the case in the living animal. In the lumbar region in particular the distal ends of the transverse processes are crushed so that they lie above the normal position. The tail is largely restored, and its length and position are somewhat conjectural. The arch of the vertebral column is that suggested by the articulations of the vertebrae, but crushing may have altered the original condition somewhat so that the arch is too pronounced.

EROSION SURFACES OF WESTERN AND CENTRAL NEW YORK

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ABSTRACT

The occurrence of two major erosion surfaces in south-central and western New York is substantiated and a correlation with those of western Pennsylvania and eastern Ohio suggested. A younger surface and a strath are found in the northern portion of western and central New York, although they are largely concealed by glacial drift. The effect of structure upon the development and preservation of these surfaces is considered.

INTRODUCTION

The publication in 1903 of an article by M. R. Campbell[1] marked the first extensive attempt to identify and correlate the erosion surfaces of southern New York with those of adjacent areas. In a recent article[2] Campbell states:

> The principal result in 1903, and the one that seems to be of far-reaching importance in unraveling the Tertiary history of the northern Appalachians, was the recognition of a partial cycle of erosion that produced on relatively weak rocks a well-developed plain, which the writer then named the Harrisburg peneplain.

This surface was traced at that time from its type locality at Harrisburg, Pennsylvania, northward across Pennsylvania into southern New York and thence southwestward into western Pennsylvania and eastern Ohio.

Fridley[3] in 1929 challenged the correlation of the erosion surface on the upland of south-central New York with the Harrisburg erosion surface as originally defined by Campbell. Ver Steeg[4] believes, however, that it is not only possible to trace the Harrisburg surface

[1] "Geographic Development of Northern Pennsylvania and Southern New York," *Geol. Soc. Amer. Bull. 14* (1903), pp. 277–96.

[2] "Chambersburg (Harrisburg) Peneplain in the Piedmont of Maryland and Pennsylvania," *Geol. Soc. Amer. Bull. 44* (1933), p. 555.

[3] H. M. Fridley, "Identification of Erosion Surfaces in South-Central New York," *Jour. Geol.*, Vol. XXXVII (1929), p. 128.

[4] K. Ver Steeg, "Some Features of Appalachian Peneplanes," *Pan-Amer. Geol.*, Vol. LIII (1930), pp. 361–64; *ibid.*, Vol. LIV (1930), pp. 17–22.

from its type locality into south-central New York but also that it may be definitely located in western Pennsylvania and eastern Ohio.

The classification and identification of erosion surfaces by Ver Steeg is essentially the same as that proposed by Campbell in 1903. The writer, however, has definitely proved from studies conducted in eastern Ohio and adjacent areas that there are three major erosion surfaces and one strath instead of the three surfaces postulated by Ver Steeg.

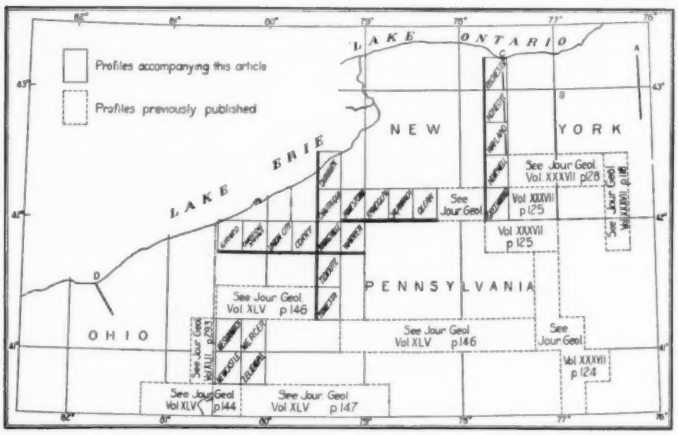

FIG. 1.—Index map, showing portions of New York, Ohio, and Pennsylvania. The heavy line in each section indicates the base line of the projected profile.

The purpose of the present article is to extend the surfaces recognized by the writer[5] in eastern Ohio and western Pennsylvania into western and central New York. The index map (Fig. 1) shows the area over which detailed studies were made.

TIONESTA-DUNKIRK SECTION

From the southern edge of the Tionesta quadrangle (Fig. 2, *top*) with average upland elevations of about 1,600 feet, there is a gradual

[5] W. S. Cole, "Identification of Erosion Surfaces in Eastern and Southern Ohio," *Jour. Geol.*, Vol. XLII (1934), pp. 285–94; "Rock Resistance and Peneplain Expression," *ibid.*, Vol. XLIII (1935), pp. 1049–62; "Development and Structural Control of Erosion Surfaces," *ibid.*, Vol. XLV (1937), pp. 141–57.

increase in elevation northward to a culminating area of 1,900+ feet on the Youngsville quadrangle. Slightly lower elevations are maintained across the Chautauqua and southern portion of the Dunkirk quadrangles. The marked escarpment which runs diagonally northeast across the Dunkirk quadrangle is shown by the rapidly descending lines from the center to the northern edge of the profile of this quadrangle.

An east-west projected profile of the Tionesta quadrangle has been previously published and described.[6] The surface represented by the upland summits of this quadrangle is the typical Allegheny erosion surface. The present profile indicates that this surface may be traced northward to the escarpment on the Dunkirk quadrangle.

GIRARD-WARREN SECTION

This profile (Fig. 2, *upper center*) should be compared to the Shenango-Tionesta profile previously published.[7] The break between the Lexington and Allegheny surfaces may be observed near the western edge of the Cambridge Springs quadrangle. This break is of the same type and magnitude as that shown on the projected profile of the Shenango quadrangle.

Eastward the Allegheny surface rises gradually. The few summit elevations of about 2,200 feet at the eastern edge of the profile of the Warren quadrangle represent the intermediate summit elevations between the typical Allegheny surface and the typical Upland surface.

As the Youngsville quadrangle has been used in the construction of the Tionesta-Dunkirk profile as well as in Girard-Warren profile, an east-west as well as a north-south profile of the Allegheny surface is presented for this quadrangle.

JAMESTOWN-OLEAN SECTION

The projected profile of these quadrangles (Fig. 2, *lower center*) connects the Girard-Warren section (Fig. 2, *upper center*) with the Woodhull-Belmont section,[8] previously published by Fridley. Two

[6] Cole, "Development and Structural Control of Erosion Surfaces," *op. cit.*, pp. 146, 149.

[7] *Ibid.*, p. 146, Fig. 3, *top*.

[8] Fridley, *op. cit.*, p. 125, Fig. 6.

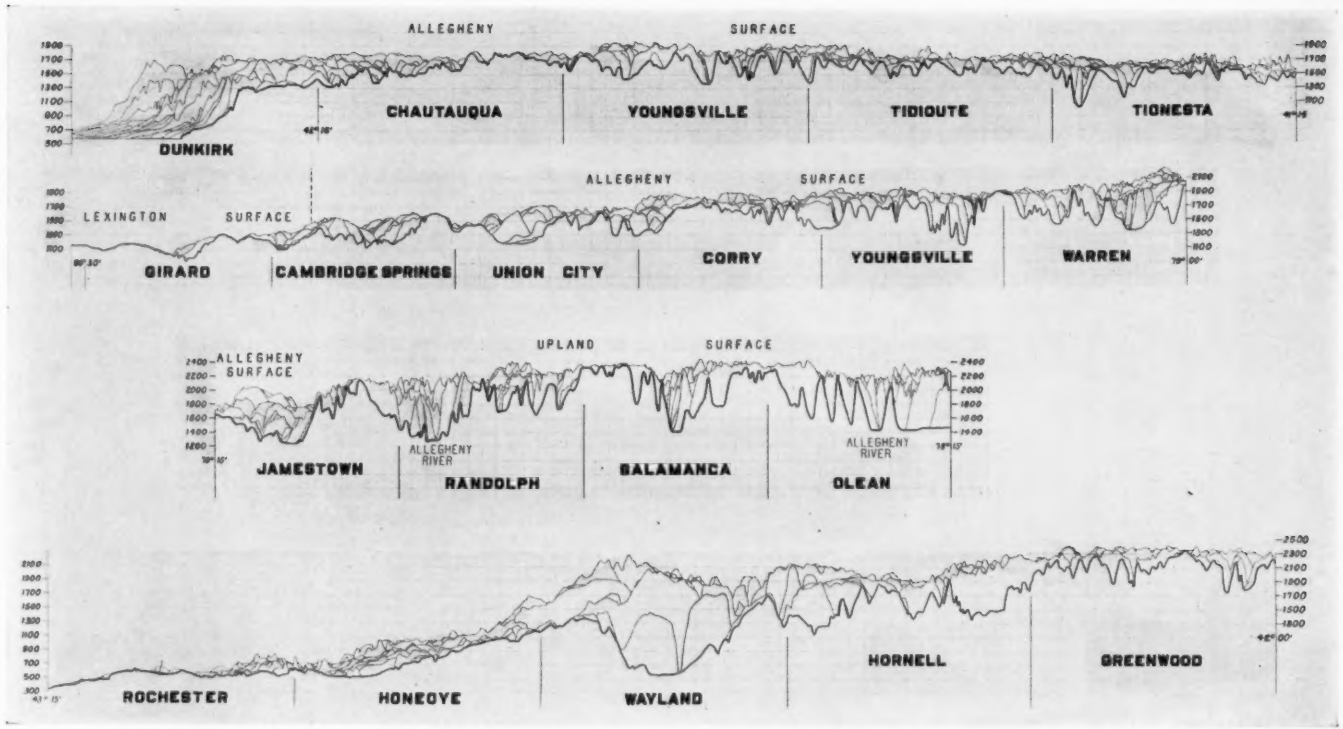

FIG. 2.—*Top:* projected profiles of the Tionesta-Dunkirk section. Base line along the western boundary of the quadrangles. *Upper center:* projected profiles of the Girard-Warren section. Base line along the southern boundary of the quadrangles. *Lower center:* projected profiles of the Jamestown-Olean section. Base line along the southern boundary of the quadrangles. *Bottom:* projected profile of the Rochester-Greenwood section. Base line along the western boundary of the quadrangles.

distinct topographic levels are shown on both of these projected profile sections, the lower surface with an altitude about 1,800 feet and the upper with an altitude of 2,300–2,400 feet.

The fact that two distinct surfaces occur in this area has been recognized previously by Campbell.[9] On the Hammondsport and Corning quadrangles just east of the area shown by the combined Jamestown-Woodhull sections, Ver Steeg[10] has identified two erosion surfaces.

The existence of two separate and distinct upland surfaces in this portion of the Appalachian plateaus seems well substantiated, but the correlation of these surfaces with those of other areas is still in doubt. Campbell correlated the upper surface with the Schooley, while Fridley identifies it as Kittatinny. Ver Steeg has combined the Kittatinny and Schooley surfaces. The lower surface was correlated by Campbell with the Harrisburg, but Fridley applies the name Schooley to this surface. Ver Steeg is in agreement with Campbell and terms the lower surface the Harrisburg.

From detailed studies in eastern and southern Ohio, and adjacent areas in Pennsylvania and West Virginia, the writer has concluded that there are three major and one minor erosion surfaces. These have been named from oldest to youngest: Upland, Allegheny, Lexington (Worthington), and Parker strath. These correlations are summarized in Table 1.

The projected profile of the Tionesta-Dunkirk quadrangles has definitely proved that the Allegheny surface may be continued from the uplands of western Pennsylvania into southwestern New York. Thus, the Harrisburg surface of Campbell and Ver Steeg and the Schooley surface of Fridley is the equivalent of the Allegheny surface of western Pennsylvania and West Virginia.

ROCHESTER-GREENWOOD SECTION

Three distinct surfaces may be recognized on this profile (Fig. 2, *bottom*). Across the Greenwood quadrangle elevations between 2,300 and 2,400 feet are maintaining the highest surface. The Greenwood

[9] "Geographic Development of Northern Pennsylvania and Southern New York," *op. cit.*, pp. 281–82.

[10] *Op. cit.*, pp. 359–61.

quadrangle was used by Fridley in the construction of the Woodhull-Belmont section. The nomenclature of this surface has been discussed under the Jamestown-Olean section. It should be correlated with the Upland surface.

Slightly lower summit elevations are observed on the profile of the Hornell and Wayland quadrangles. These are classified as intermediate elevations which have been reduced below the elevations of

TABLE 1

Central and Western New York				Western Pennsylvania and Eastern Ohio
Campbell (1903)	Fridley (1929)	Ver Steeg (1930)	Cole (this article)	Cole (1934–35–37)
Schooley	Kittatinny	Schooley (Kittatinny)	Upland	Upland
Harrisburg	Schooley	Harrisburg	Allegheny	Allegheny
Not studied	Mine Ridge	Not studied	Niagara	Lexington (Worthington)
	Three erosion surfaces below Mine Ridge recognized north of Ithaca, N.Y., but not correlated			
			Floors of preglacial valleys	Parker strath

the Upland surface but have not been sufficiently lowered to be classified as Allegheny. This relationship is the natural and expected one in the border zone between two erosional surfaces.

The profile of the Hornell and Wayland quadrangles as well as the contour map of summit elevations shows that the Allegheny surface is fairly well developed surrounding these intermediate summit areas.

A lower surface is shown on the Rochester and northern portion of the Honeoye quadrangles. The relationship of this surface to the

Allegheny definitely establishes its correlation with the Lexington (Worthington) surface of eastern Ohio. If it is found desirable to designate this surface by a local name, the term Niagara erosion surface is suggested.

Below this surface is found the equivalent of the Parker strath of western Pennsylvania and eastern Ohio, represented by the floors of preglacial valleys whose streams were actively engaged in the removal of this surface (Niagara) at the time of the first advance of the Pleistocene ice. Unfortunately the strath stage of this area either has been destroyed by the advancing ice over deepening the valley floors or has been concealed by thick ice deposits. Ver Steeg[11] and the writer[12] have considered that the buried preglacial valleys of western Ohio represent Parker time, and thus the erosional history of this area is identical with that of the Till Plain and Eastern Lake Plain sections of Ohio. In this connection, however, it must be stated that there is grave danger of confusing valleys developed during one or more interglacial stages with the preglacial valleys. It is the belief of the writer that this has commonly occurred.

Comparison of this projected profile with the structure section (Fig. 7, *C*) at once demonstrates that the flagstones and sandstones of the Upper Devonian are one of the dominant controls in the preservation of both the Allegheny and the Upland surfaces of this area.

COMPARISON OF THE CONTOUR AND AREAL GEOLOGIC MAPS

New York.—A comparison of the contour map of summit elevations of western and central New York (Fig. 3) with the areal geologic map (Fig. 4) indicates the control which certain formations exert in maintaining the upland areas. The trend of the contours is identical with the geologic boundaries.

Except in the area east of Cayuga Lake, rocks of Chemung age cap the upland summits. East of Cayuga Lake, however, the highest summits are north of the Chemung boundary and are capped by

[11] "The Buried Topography of North-Central Ohio and Its Origin," *Jour. Geol.*, Vol. XLII (1934), p. 610.

[12] "Identification of Erosion Surfaces in Eastern and Southern Ohio," *op. cit.*, p. 291.

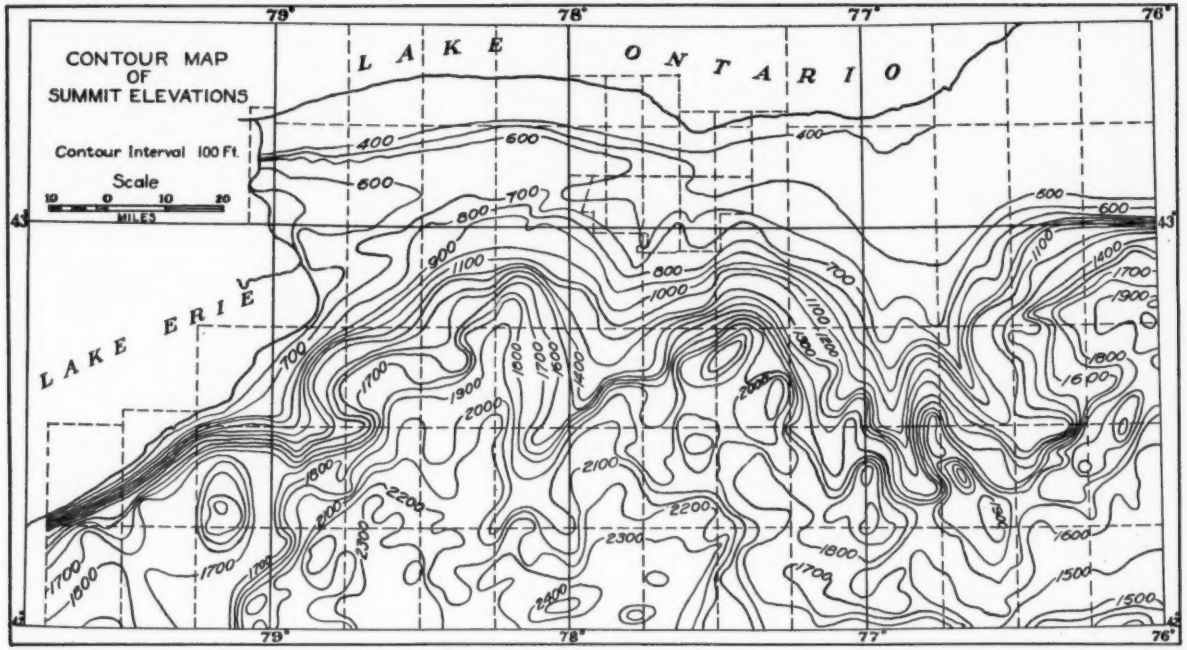

FIG. 3.—Contour map of western and central New York, drawn from hilltop elevations, to show the major features and areal expression of the erosion surfaces. The dotted lines indicate the position of the quadrangles.

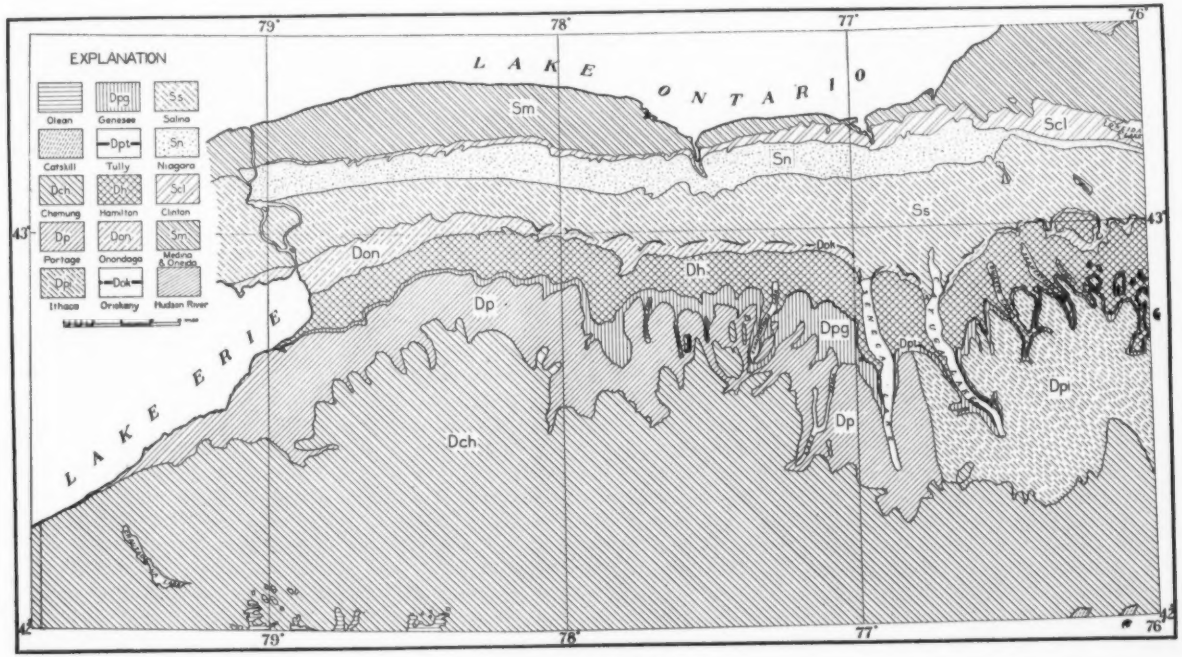

FIG. 4.—Areal geologic map

rocks of Portage age.[13] Keith[14] in 1923 published a generalized map which indicated a syncline approximately through the present basins of Cayuga and Seneca lakes. Von Engeln[15] considers that the broad, low synclinal depression described above has been an important factor in the development of this part of the "central New York lowland," which includes the middle and northern sections of the region here described.

That same year (1932) Wedel[16] published a map, accompanied by a detailed description, illustrating the structure of this region. Wedel clearly demonstrates that if a broad north-south syncline is present, its axis lies near the Dansville meridian.[17] If this map had been available to von Engeln, without doubt he would have assigned other reasons for the development of Cayuga-Seneca lowland re-entrant.

The explanation of the localization of the major troughs of Cayuga and Seneca is associated with the change of strike[18] just west of Seneca Lake as well as lithic and thickness changes of the formations in this area.

Note should be made of other re-entrants which penetrate the Chemung in various places, although marked uplands partially surround them. These are interpreted as major erosional lowlands developed during Allegheny time.

Ohio and Pennsylvania.—The contours of the Ohio portion of this map (Fig. 5) show the northward continuation of the Mississippian cuesta[19] which trends northeast toward the Pennsylvanian boundary. The escarpment slope of the cuesta continues into Pennsylvania and New York, but the dip slope is lost. Comparison of these con-

[13] Recently a new classification has been proposed for the Upper Devonian of New York (G. H. Chadwick, "Chemung Is Portage," *Geol. Soc. Amer. Bull. 46* [1935], pp. 343–54). For the purposes of this paper the nomenclature used on the various maps is employed as problems of stratigraphic correlation are not involved.

[14] A. Keith, "Outlines of Appalachian Structure," *Geol. Soc. Amer. Bull. 34* (1923), p. 309, Pl. IV.

[15] O. D. von Engeln, "The Finger Lake Region," in "Paleozoic Stratigraphy of New York," *XVIth International Geol. Cong. Guidebook 4* (1932), pp. 49–51.

[16] A. A. Wedel, "Geologic Structure of the Devonian Strata of South-Central New York," *New York State Mus. Bull. 294* (1932), pp. 1–74, 1 map.

[17] *Ibid.*, p. 17. [18] *Ibid.*, Pl. I.

[19] Cole, "Rock Resistance and Peneplain Expression," *op. cit.*, p. 1052, Fig. 2; pp. 1056–59.

tour lines with the formational boundaries indicates the parallelism between the trend of rock units and the topography.

Strikingly, however, the contours of the remainder of both the Ohio and the Pennsylvania portions of the map do not parallel the formational boundaries. With the exception of those contours in northwestern Pennsylvania which parallel the shore of Lake Erie, the remainder of the contours in this portion of the map trend at

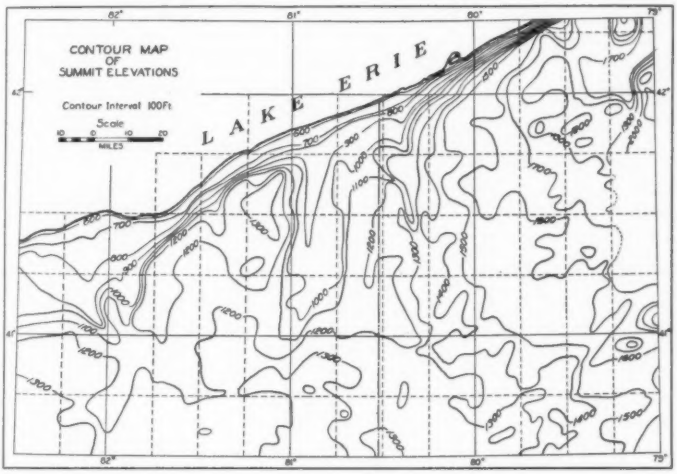

FIG. 5.—Contour map of a portion of northwestern Pennsylvania and northeastern Ohio, drawn from hilltop elevations to show the major features and areal expression of the erosion surfaces.

nearly right angles to the formational boundaries. The origin of the marked low area which extends northwestward into Ohio from the southwestern corner of the Pennsylvania portion of the map has been previously discussed.[20] This lowland is a re-entrant of both the Lexington and the Parker strath surfaces into the Allegheny, which is maintained by uplands on either side. Inasmuch as this lowland trends nearly at right angles to the regional strike, it cuts across formations of varying resistances, including those which are controlling the Mississippian cuesta.

[20] Cole, "Development and Structural Control of Erosion Surfaces," *op. cit.*, p. 154.

With the exception of the escarpment paralleling Lake Erie, the contours of summit elevations in northeastern Ohio and northwestern Pennsylvania do not exhibit structural control. Instead they reflect erosional conditions, particularly of an area where a lower surface was being actively etched from an older and higher surface along a major stream.

THE MODEL

In order that the concepts which are presented in cross-sectional view by projected profiles and in plan by the maps of summit eleva-

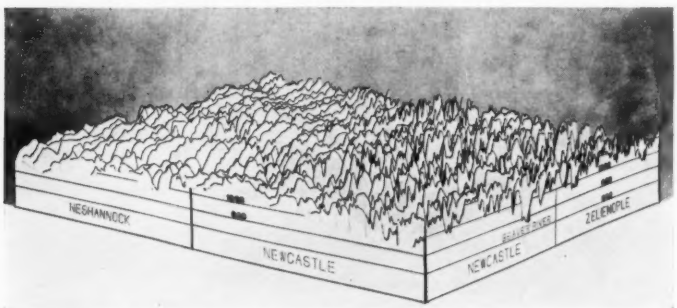

FIG. 6.—Photograph of a model constructed by using east-west profiles at 1-inch intervals of four quadrangles of western Pennsylvania. (Mercer quadrangle lies east of the Neshannock.)

tions might be checked, a three-dimensional model of a selected area was constructed (Fig. 6). The area chosen was covered by four topographic maps: Newcastle; Zelienople; Neshannock; and Mercer, Pennsylvania.

A series of east-west profiles was drawn across these maps at intervals of 1 inch. The horizontal scale is that of the maps, 1 inch to approximately 1 mile, and the vertical scale is 1 inch to 200 feet. The vertical exaggeration is the same as that of the projected profiles. The profiles were transferred to cardboard and cut out. The cardboard strips were correctly oriented in a frame.

While projected profiles were not made of these four quadrangles in order to avoid duplication, they have been constructed for quad-

rangles immediately north[21] and south.[22] As the quadrangles for which projected profiles were made have the same erosional history as those involved in the model, it was thought that the model should exhibit the same features as the adjacent projected profiles.

Detailed study of the model proves that the same concepts are obtained from it as are gained by the examination of the projected profiles and maps of summit elevations. The projected profiles or the maps of summit elevations used singly will not give so accurate a concept of the relationships as will a model of this type. However, the combined projected-profile and summit-elevation map will present exactly the same concept of the erosion surfaces as will the model. Inasmuch as fifty hours were utilized in the construction of the model, while both projected profiles and a map of summit elevations of the same area could be constructed in ten hours at a maximum, it is obvious that the latter system is more practical for regional studies of erosion surfaces.

The area covered by the model represents the eastern portion of a Lexington re-entrant into the Allegheny surface. While the major stream of the area is the south-flowing Beaver River formed by the juncture of the Mahoning and Shenango rivers in the north-central portion of the Newcastle quadrangle, the slope of the upland as outlined by the hilltops is toward the northwest corner of the model. This slope is portrayed also on the map of summit elevations.

Detailed study of the model does substantiate the presence of the Parker strath below the upland and above the present floor of the Beaver River. This is the only additional fact brought out by the model. The streams of Lexington and Parker-strath times were north-flowing as the general slope of the upland indicates. Reversal of drainage followed the first advance of the Pleistocene ice.

Detailed geologic reports are available for the Newcastle[23] and Zelienople[24] quadrangles. The hills of the area covered by these two

[21] *Ibid.*, p. 146, Fig. 3, *top*, Shenango and Stoneboro quadrangles.

[22] *Ibid.*, p. 144, Fig. 2, *middle*, Beaver quadrangle; and p. 147, Fig. 4, *top*, Sewickley quadrangle.

[23] F. W. De Wolf, "Geology and Mineral Resources of the New Castle Quadrangle," *Top. and Geol. Atlas of Penn. No. 5* (1929), pp. 1–238, 9 pls.

[24] G. B. Richardson, "Geology and Mineral Resources of the Butler and Zelienople Quadrangles, Pennsylvania," *U.S. Geol. Survey Bull. 873* (1936), pp. 1–93, 8 pls.

quadrangles are capped by rocks of Allegheny or Conemaugh age. Rock resistance is not the dominant control in shaping the upland features of these quadrangles. The configuration of the upland slope is entirely a function of the north-flowing, preglacial stream systems of the area.

STRUCTURAL SECTIONS

These sections (Fig. 7) illustrate the same features that may be noted by comparing the maps of summit elevations with areal geologic maps. The formational names used in this article are those employed by the various geologists who did the mapping.

The Syracuse-Tully section (Fig. 7, *top*) shows that the major escarpment in this area is controlled by rocks of Portage age (Sherburne and Ithaca). A minor escarpment below the major one of this section is controlled by the Onondaga limestone. If the drift were removed from the area north of Syracuse, the Lockport dolomite would, without doubt, stand out as a distinct relief feature, as it does at the present time in the Niagara Falls region.

The Geneva-Elmira (Fig. 7, *upper center*) and Rochester-Wayland (Fig. 7, *lower center*) sections demonstrate that the highest hills are capped with Chemung beds. The Portage beds which were capping the hilltops on the Syracuse-Tully section appear in the escarpment face.

The westernmost section (Fig. 7, *bottom*) was constructed across the Mississippian cuesta just west of the city of Cleveland, Ohio. The Sharon conglomerate caps the hills in this area. A minor shelf is controlled by the Berea sandstone.

THE MAJOR ESCARPMENT

The major escarpment (Fig. 8) which is distinctly shown by the maps of summit elevations may be traced continuously from central New York across the northwestern corner of Pennsylvania into northeastern Ohio, and thence through central Ohio into Kentucky.

This escarpment is maintained from east to west by rocks of different ages, but has had the name Portage[25] applied to it through-

[25] H. P. Cushing, F. Leverett, and F. R. Van Horn, "Geology and Mineral Resources of the Cleveland District, Ohio," *U.S. Geol. Surv. Bull. 818* (1931), p. 3.

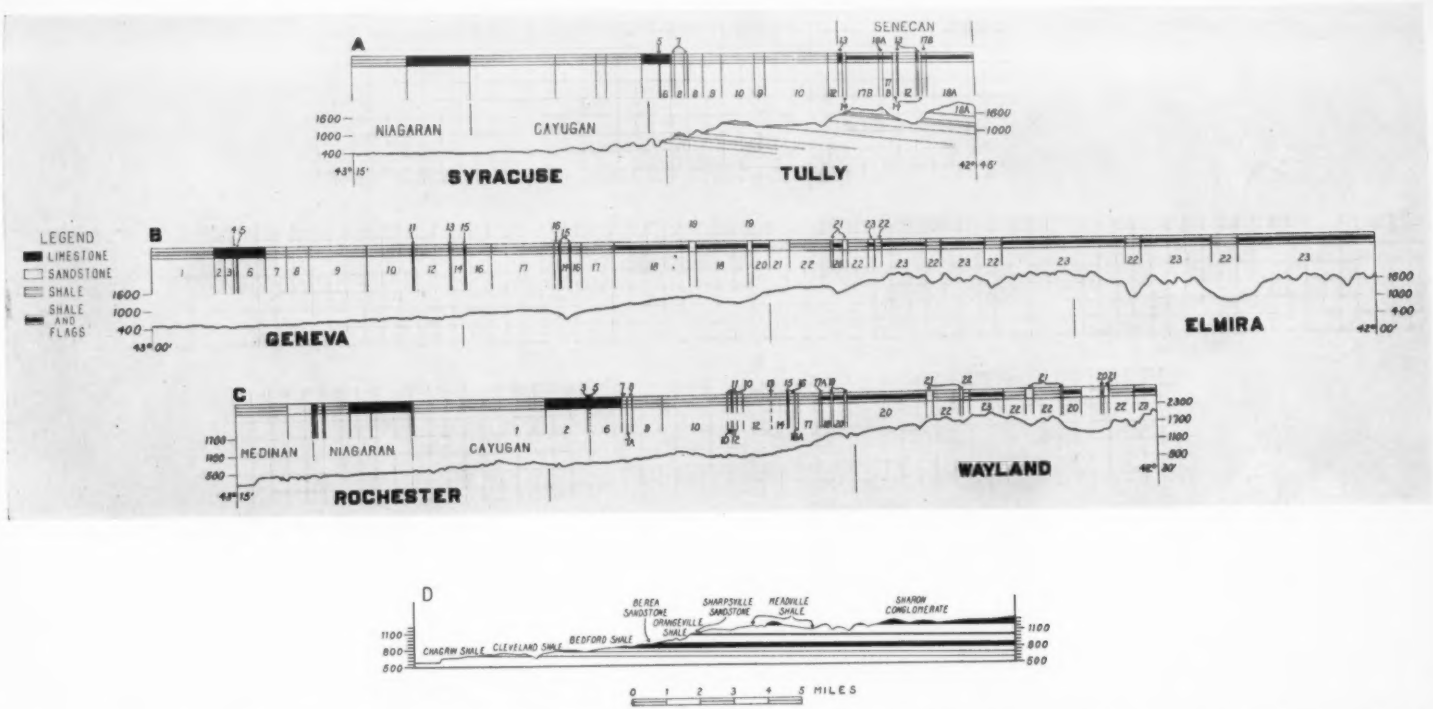

FIG. 7.—Structural and lithic sections. General orientation north-south. *A*, *B*, *C*, from New York; *D*, from Ohio. The formations represented are: (1) Camillus shale, (2) Bertie waterlime, (3) Cobleskill limestone, (4) Rondout waterlime, (5) Oriskany sandstone, (6) Onondaga limestone, (7) Marcellus shale, (7*A*) Stafford limestone, (8) Cardiff shale, (9) Skaneateles shale, (10) Ludlowville shale, (11) Tichenor limestone, (12) Moscow shale, (13) Tully limestone, (14) Genessee shale, (15) Genundewa limestone, (16) West River shale, (16*A*) Middlesex shale, (17) Cashaqua shale, (17*A*) Rhinestreet shale, (17*B*) Sherburne flags, (18) Hatch shale and flags, (18*A*) Ithaca shale and flags, (19) Grimes sandstone, (20) West Hill shale and flags, (21) High Point sandstone, (22) Wiscoy shale, (23) Chemung shale and sandstone.

out its length. This implies, incorrectly, that Portage beds control the escarpment throughout its length. It seems desirable to name that portion of the escarpment through Ohio and Kentucky the Mississippian escarpment, as various formations of that system exert the dominant control upon its configuration.

FIG. 8.—View of the Portage escarpment from the road corner about 0.7 of a mile northwest of Richmond Mills, Honeoye quadrangle, N.Y.

SUMMARY

Three major erosion surfaces occur in the area studied, with a fourth or strath stage largely destroyed or concealed in the New York section. The Allegheny and Upland surfaces may be traced directly from Pennsylvania into New York.

Structure and rock resistance are the major controls not only of the configuration of the surfaces but in their preservation. That the present terrain is not the result of a single cycle of erosion may be definitely proved by the interfingering of younger surfaces with older, as well as by numerous upland slopes which are in accord with restored drainage systems.

No attempt was made to unravel the nomenclature of the erosion surfaces of eastern Pennsylvania in this article. This analysis appears to be the next logical step in the interpretation of the erosion surfaces of the northern Appalachian region.

THE ALEXANDRIAN SERIES IN IOWA

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ABSTRACT

Since 1914, when Savage named the Waucoma formation of northeastern Iowa and referred it to the Alexandrian series, there has been a question of whether or not there are Lower Silurian strata in Iowa, and if so, what part of that series is represented. As a result of the present investigation it is proved beyond reasonable doubt that Alexandrian beds outcrop in several counties in northeastern Iowa. About 70 feet of beds formerly referred to the lower part of the Hopkinton formation of the Niagaran series are placed in the Alexandrian series. The Illinois formation names, Edgewood and Kankakee, were extended to the Iowa sections and the writer has tried to define them so that they are as nearly as possible the equivalents of those formations in Illinois. Typically, the Edgewood formation consists of about 30 to 40 feet of thin- to medium-bedded dolomite, and the overlying Kankakee formation of about the same thickness of dolomite with numerous bands of chert. The available evidence seems to indicate that these two formations are conformable with each other and that the Kankakee is conformable with the Niagaran series above.

INTRODUCTION

Since 1914, when Savage[1] named the Waucoma formation of northeastern Iowa and referred it to the Alexandrian series, there has been a question as to the formations, distribution, and relations of the strata that belong in the Alexandrian series in Iowa. The present investigation was undertaken in an attempt to answer these questions.

The field work was done during the latter part of the summer of 1934. The study of the samples, fossils, and literature was completed early in 1935 and the data and conclusions incorporated in a Master's dissertation at the University of Iowa. Some of the results have since been used by the Kansas Geological Society and the Iowa Geological Survey. Additional field work and study have necessitated some revision, but the general conclusions remain the same.

The strata in question outcrop in eight counties in northeastern Iowa in a northwest-southeast belt from southern Winneshiek and central Bremer counties to east-central Clinton County (Fig. 1).

[1] T. E. Savage, "The Relations of the Alexandrian Series in Illinois and Missouri to the Silurian Section of Iowa," *Amer. Jour. Sci.*, Vol. XXXVIII (1914), pp. 34–36.

Exposures are confined to road cuts, quarries, and a few of the deeper-stream valleys. Especially in the western part of the area the exposures are scarce because of the thickness of the glacial drift. These beds dip gently to the southwest as do all the Paleozoics of eastern Iowa.

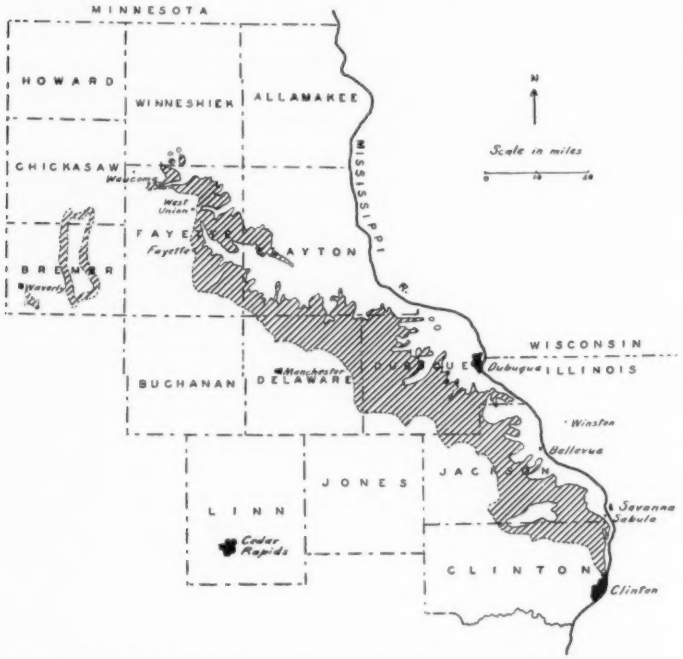

FIG. 1.—Areal geologic map showing the distribution of the Alexandrian series in Iowa.

The writer wishes especially to thank Professor A. C. Trowbridge for generous advice on problems that arose during the progress of the investigation. Professor T. E. Savage of the University of Illinois, Mr. L. E. Workman of the Illinois Geological Survey, Professor A. K. Miller and Mr. W. M. Furnish of the University of Iowa also assisted the writer.

RÉSUMÉ OF THE LITERATURE

Strata of post-Richmond and pre-Clinton age were first recognized in the Mississippi Valley in 1908 when Savage[2] proposed the term Alexandrian for some beds in Alexander County, Illinois, which lie between the Cincinnatian and the Clinton. At that time he included only the Cape Girardeau limestone in the Alexandrian series. The next year Savage[3] took the top beds out of this formation and called them the Edgewood formation with the type section at Edgewood in Pike County, Missouri. After Schuchert[4] placed the Brassfield ("Ohio Clinton"), which Foerste[5] had named and considered to be of Clinton age, below the Clinton of New York, Savage[6] placed the Sexton Creek (now Kankakee), a correlative of the Brassfield, in the Alexandrian series. In 1914, Savage,[7] after studying the Alexandrian strata in northwestern Illinois and northeastern Iowa, gave two new formation names to what he thought were correlatives of the Edgewood and Sexton Creek formations of southern Illinois and Missouri. The Winston formation was named for beds exposed at the station of Winston on the Chicago Great Western Railroad, 6 miles south of Galena in Jo Daviess County, Illinois. It was thought to have been deposited during Edgewood time but in a different basin. The Waucoma formation was named for beds outcropping along the banks of the Little Turkey River near Waucoma in Fayette County, Iowa, that were thought to represent deposition in a northern embayment during Sexton Creek time. In 1916 Sav-

[2] "On the Lower Paleozoic Stratigraphy of Southwestern Illinois," *Amer. Jour. Sci.*, Vol. XXV (1908), pp. 433–34.

[3] "The Ordovician and Silurian Formations in Alexander County, Illinois," *Amer. Jour. Sci.*, Vol. XXVIII (1909), pp. 516–18.

[4] Charles Schuchert, "Paleography of North America," *Geol. Soc. Amer. Bull. 20* (1909), pp. 533–38.

[5] A. F. Foerste, "The Silurian, Devonian and Irvine Formations of East-Central Kentucky," *Ky. Geol. Surv. Bull. 7* (1906), p. 18.

[6] "Alexandrian Series in Missouri and Illinois," *Geol. Soc. Amer. Bull. 24* (1913), pp. 370–75.

[7] "The Relations of the Alexandrian Series in Illinois and Missouri to the Silurian Section of Iowa," *op. cit.*, pp. 34–36.

age[8] proposed the term Kankakee limestone for exposures along the Kankakee River, 5 miles south of Rickey in Will County, Illinois, and considered them a northern embayment equivalent of the Sexton Creek formation. In this same paper the base of the Kankakee was defined as occurring at the *Platymerella manniensis* zone. In 1926 Savage[9] placed the top boundary of the Kankakee just above the *Stricklandinia pyriformis* zone where a smooth but pitted plane was thought to represent an unconformity.

During the early part of the present century various geologists prepared for the Iowa Geological Survey reports on the geology of the northeastern counties of Iowa, and in all of these the beds that are now recognized as Alexandrian in age were referred to the Niagaran. In 1914 Savage, as already stated, named the Winston and Waucoma formations of the Alexandrian and described part of their general outcrop area in Iowa. In 1926 he listed Edgewood and Waucoma on a general correlation chart as the Alexandrian formations of Iowa (Table 1). The following year the Iowa Geological Survey placed the Waucoma formation in the general geological column of the state.[10] In 1935 Sutton[11] wrote a brief discussion of the Alexandrian series in the Upper Mississippi Valley, in which he used the terms Edgewood and Kankakee as formation names in Iowa to replace Winston and Waucoma, respectively, as Savage had done for northwestern Illinois.

FAUNA OF THE ALEXANDRIAN

The Alexandrian beds contain a rather meager fauna. They were referred to in some of the old Iowa county reports as the unfossiliferous basal beds of the Niagaran. However, in general, the Alexandrian fossils are well preserved, in contrast to the overlying Niagaran fossils, most of which occur as rough molds. Many of the species are new but in all cases they appear to be closely related to described forms.

[8] "Alexandrian Rocks of Northeastern Illinois and Eastern Wisconsin," *Geol. Soc. Amer. Bull. 27* (1916), pp. 315–16.

[9] "Silurian Rocks of Illinois," *Geol. Soc. Amer. Bull. 37* (1926), pp. 518–19.

[10] G. F. Kay, "Rock Resources of Iowa," *Iowa Geol. Surv.*, Vol. XXXII (1927), p. 52.

[11] A. H. Sutton, "Stratigraphy of the Silurian System of the Upper Mississippi Valley," *Kansas Geol. Soc., Ninth Ann. Field Conf. Guide Book* (1935), pp. 275–76.

STRATIGRAPHY OF THE ALEXANDRIAN SERIES

THE EDGEWOOD FORMATION

Fauna.—The best index fossil to the Edgewood in Iowa is the common brachiopod *Plectatrypa praemarginalis*. *Wattsella edgewoodensis* is supposed to be a strictly Edgewood species but a variety of it was found in what is classified as the lower part of the Kankakee. *Platystrophia daytonensis* is common but it also occurs sparingly in the Kankakee. *Calymene vogdesi* marks a definite zone near the top of the Edgewood. Other rarer species were found that are confined to the Edgewood and these prove the presence of Edgewood strata in Iowa.

TABLE 1

CHART SHOWING HISTORY OF SAVAGE'S USE OF FORMATIONAL NAMES WITHIN THE ALEXANDRIAN

SW. Illinois E. Missouri 1908	SW. Illinois E. Missouri 1909–10	Illinois Missouri 1913	Iowa NW. Illinois 1914
		Sexton Creek	Waucoma
Cape Girardeau	Edgewood	Essex	Winston
		Edgewood	
	Girardeau	Girardeau	

S. Illinois NE. Illinois E. Missouri 1914	NE. Illinois 1916 NE. Illinois NW. Illinois 1926	SW. Illinois 1926	Iowa 1926
Sexton Creek	Kankakee	Sexton Creek	Waucoma
Edgewood	Edgewood	Edgewood	Edgewood
Girardeau		Girardeau	
		Orchard Creek	

Lithology.—The Edgewood in most places is a thin-bedded, yellow-gray, siliceous dolomite. White chert nodules occur indiscriminately throughout the formation. It varies in bedding to be-

TABLE 2

TABLE SHOWING THE FOSSILS OF THE EDGEWOOD AND KANKAKEE FORMATIONS OF IOWA

	Edgewood	Kankakee
Zaphrentis subregularis Savage	x	...
Zaphrentis sp.	x	x
Favosites favosus (Goldfuss)	x	x
Favosites niagarensis Hall	x	x
Aulopora sp.	x	x
Syringopora(?) sp.	...	x
Halysites catenularia (Linnaeus)	x	x
Heliolites sp.	...	x
Clathrodictyon(?) sp.	x	x
Helopora sp.	x	...
Nematopora sp.	x	x
Lingula cuneata Conrad(?)	x	...
Orthis aff. *O. laurentia* Billings	...	x
Dolerorthis flabellites (Foerste)	x	x
**Dolerorthis interplicata* (Foerste)(?)	x	...
Platystrophia daytonensis (Foerste)	x	x
Platystrophia reversata (Foerste)	x	x
Wattsella edgewoodensis (Savage)	x	...
Wattsella edgewoodensis var. (Savage)	...	x
Rhipidomella hybrida (Sowerby)(?)	x	x
Sowerbyella transversalis (Whalenberg)	x	x
Sowerbyella transversalis elegantula (Hall)(?)	x	x
Leptaena rhomboidalis (Wilkens)	x	x
Strophonella(?) cf. *S. daytonensis* (Foerste)	...	x
Schuchertella cf. *S. subplana* (Conrad)	x	...
Schuchertella aff. *S. curvistriata* Savage	x	...
Schuchertella roemeri (Foerste)(?)	...	x
Stricklandinia aff. *S. pyriformis* Savage	...	x
Rhynchotreta intermedia Savage	x	...
Rhynchotreta parva Savage	x	...
Camerotoechia whitei praecursor (Foerste)	...	x
Atrypa reticularis (Linnaeus)	...	x
Atrypa nodostriata Hall	...	x
**Atrypa putilla* (Hall and Clarke)	x	...
Plectatrypa aff. *P. marginalis* (Dalman)	...	x
Plectatrypa praemarginalis (Savage)	x	...
Plectatrypa tubulistriata (Savage)(?)	x	...
Bellerophon(?) sp.	x	...

* Found only in northwestern Illinois.

TABLE 2—*Continued*

	Edgewood	Kankakee
**Lophospira fasciata* Savage(?)	x	...
Hormotoma cf. *H. subulata* (Conrad)	...	x
Diaphorostoma cf. *D. illinoisensis* Savage	x	...
Cyclonema daytonensis Foerste	x	...
Kionoceras sp.	x	...
Huronia obliqua Stokes	x	...
Huronia vertebralis Stokes	x	...
Leperditia sp.	x	x
Illaneus aff. *I. imperator* Hall	x	...
Illaneus daytonensis Hall and Whitfield(?)	x	x
Illaneus slocomi Walter(?)	x	...
Odontopleura cf. *O. ortoni* (Foerste)	x	...
Encrinurus aff. *E. thresheri* Foerste	x	x
Calymene vogdesi Foerste	x	...
Metapolichas cf. *M. breviceps clintonensis* (Foerste)	x	...

come thick-bedded or massive in places. In color it varies from buff to gray and in some places it is bluish-gray immediately above the Maquoketa shale. In some localities the dolomite is porous or even earthy. In northern Fayette County around West Union it is a light-gray limestone. Shaly partings at the bedding planes occur in some localities

Thickness.—The Edgewood is characteristically variable in thickness because of the disconformity at its base. It is thick where the Maquoketa is thin, and vice versa. However, the Edgewood does tend to thicken toward the north. From 30 to 40 feet is its average thickness but it varies from less than 5 to 100 feet.

Correlation with the Winston.—The belief that the Edgewood is the equivalent of the Winston and belongs in the same sedimentary province was put forth by Savage[12] when he applied the term Edgewood to strata formerly called Winston in northwestern Illinois. If this is true Edgewood is the correct term to use, and since it is in good usage for the northwestern Illinois strata, the writer applies it to the corresponding Iowa strata.

Correlation with the Waucoma.—The writer places the beds found at the type section of the Waucoma in the Edgewood formation on

[12] "Silurian Rocks of Illinois," *op. cit.*, pp. 528–29.

the basis of paleontology, lithology, and normal stratigraphic sequence. *Dolerorthis flabellites* and *W. edgewoodensis* were the only identifiable fossils found and they indicate its Edgewood age. The beds contain no chert and nowhere in Iowa has the Kankakee been seen without some chert bands. The Edgewood is present in surrounding exposures and normally would be expected here. Thus, the terms Waucoma and Winston were really applied to the same strata that carry the present designation of Edgewood.

Stratigraphic relations.—Existing evidence indicates a definite time break between the deposition of the Maquoketa and Edgewood. The paleontological break is well shown by the fact that out of 37 genera known to occur in the Alexandrian of Iowa only 5 of the 34 genera that Ladd[13] lists from the upper Maquoketa Brainard shale are the same, and of the 20 genera listed from the middle Maquoketa Fort Atkinson dolomite only one, *Lingula*, continues up into the Silurian. There is commonly a sharp lithologic break from the blue-green clay shale of the Maquoketa to the dolomite of the Edgewood, although at Bellevue about 25 feet of blue-gray dolomitic shale with a blocky fracture cannot definitely be assigned to either formation. Their disconformable relations are shown by the fact that locally the Edgewood is thick where the Maquoketa is thin, and vice versa. This disconformity is well shown at Savanna, Illinois, where the Edgewood thickens from 4 to 12 feet in a horizontal distance of 50 feet.

The line between the Edgewood and the overlying Kankakee is much more difficult to locate. The defining fossil, *P. manniensis*, was not found in Iowa or northwestern Illinois. In most cases the contact was defined on the basis of lithology because normally there is a change from noncherty to cherty beds at approximately the horizon where the line should be drawn. This may differ slightly from where Savage draws the line in Illinois, but it is considered the only practical division because in any one exposure it is difficult to find fossils restricted to either formation near the contact.

The Edgewood appears to grade conformably into the Kankakee

[13] H. S. Ladd, "The Stratigraphy and Paleontology of the Maquoketa Shale of Iowa," *Iowa Geol. Surv.*, Vol. XXXIV (1929), pp. 391–95.

above. There is evidence of some change of conditions that may represent unconformity at Savanna, Illinois, where a deeply pitted, polished surface occurs at the top of the Edgewood. Whether this is due to submarine or to subaerial erosion, or to some organic agency, it at least represents no more than a diastem because of the almost positive evidence of conformity found in all Iowa sections.

THE KANKAKEE FORMATION

Fauna.—The Kankakee has been defined on the basis of faunal zones. A zone of *P. manniensis* marks the base and the *S. pyriformis* zone occurs near the top of the Kankakee. Both of these zones are well developed in northeastern Illinois, but *P. manniensis* has not been found in Iowa or northwestern Illinois. *S. pyriformis* has been reported from northwestern Illinois, and the writer found this zone at one place, at Sabula, Iowa. A zone of *Atrypa*, mostly *A. reticularis*, occurs 6 feet below the top of the Kankakee in many places. Other rarer species were found that are restricted to the Kankakee and these prove the presence of strata of Kankakee age in Iowa.

Lithology.—The Kankakee is characteristically a yellow dolomite interbedded with bands of white chert from 1 to 6 inches thick. In some places the chert bands make up nearly as much of the bulk of the formation as the dolomite. In general, the chert bands are persistent horizontally and vertically. In no outcrop in Iowa did the writer find undoubted Kankakee that did not contain some chert bands. This formation weathers to a reddish, oxidized material containing broken chert fragments.

Thickness.—Locally the Kankakee varies little in thickness, but it seems to thicken to the south as the Edgewood thins. Seventeen feet are present at West Union in Fayette County, and 60 feet are exposed at Clinton in Clinton County. The average thickness is about 30 to 40 feet.

Stratigraphic relations.—The Kankakee rests conformably on the Edgewood below. At no place in Iowa is it known to rest directly on the Maquoketa or on any older formation. It differs from the Edgewood only on the basis of fossils and the included chert bands. Even so, many species are common to both formations, and even the

Edgewood contains nodular or nodular-banded chert in some localities.

The chert bands of the Kankakee stop somewhat abruptly at the top and the Hopkinton above consists of massive, buff dolomite. Because of the lithologic change at this line and the presence of the zone of *Stricklandinia* just below it, the writer believes that this is the equivalent of the formational boundary drawn in eastern Illinois.

Savage[14] correlated the Joliet and Hopkinton with the Lockport, and Sutton[15] correlated the Joliet and part of the Hopkinton with the upper part of the Clinton of New York. In either case this necessitated an unconformity at the top of the Alexandrian because some beds of Clinton age would be missing. However the writer, believing that the Silurian in Iowa and northwestern Illinois are a conformable sequence, would correlate the lower part of the Joliet and Hopkinton with the Clinton.

All the Illinois writers including Athy,[16] Fisher,[17] and Savage[18] believe that an unconformity is present between the Kankakee and Joliet in northeastern Illinois. The evidence for this is a peculiarly planed and pitted surface that is supposed to mark the contact of all outcrops exposed in the Herscher, Wilmington, and Joliet quadrangles. The writer has found no such persistent surface in Iowa. In fact, comparable surfaces are present at several different horizons in both the Edgewood and Kankakee.

Except for a thin clay seam at the top of the Kankakee at Sabula, Iowa, and Savanna, Illinois—the same kind of seam is also present at other horizons in the Hopkinton—the Kankakee grades conformably into the Hopkinton. At Farley, Iowa, where an excellent section of the Edgewood, Kankakee, and Hopkinton is exposed, the upper 6 feet of the Kankakee is broken up, weathered, and inter-

[14] "Silurian Rocks of Illinois," *op. cit.*, p. 533.

[15] *Op. cit.*, pp. 268–69.

[16] L. F. Athy, "Geology and Mineral Resources of the Herscher Quadrangle," *Ill. Geol. Surv. Bull. 55* (1928), p. 45.

[17] D. J. Fisher, "Geology and Mineral Resources of the Joliet Quadrangle," *Ill. Geol. Surv. Bull. 51* (1925), pp. 31–32.

[18] "Silurian Rocks of Illinois," *op. cit.*, pp. 518–19.

spersed with clay, indicating that this stood near the surface of the water. However, no significant uplift would be necessary to bring part of the Kankakee above water, as the whole Silurian sequence must have been deposited in shallow water. Since no evidence of this uplift is found in surrounding exposures, and this general area is in the vicinity of a local structural high, and little, if any, of the Kankakee is missing here, the re-working of the Kankakee is believed to represent a local diastem and not an important unconformity.

REVIEWS

Australian Pre-Cambrian Fossils: A Memoir of the Late Pre-Cambrian Fossils from the Adelaide Series, South Australia. By SIR T. W. EDGEWORTH DAVID and R. J. TILLYARD. Sydney: Angus & Robertson, 1936. Pp. 122; pls. 13. *7s 6d.*

The late Sir T. W. Edgeworth David ranked high among the energetic and enthusiastic international group of searchers for the primitive ancestors of the early Paleozoic invertebrates. With keen insight, unwearying patience, and unflagging determination, Professor David continued to comb the pre-Cambrian sediments of South Australia almost to the time of his death in 1934. One of his early publications in connection with this problem concerned the remains of small crustacea in the rocks of Reynella, near Adelaide, but in 1922, when this paper was issued, there was a question as to whether the rocks were actually pre-Cambrian in age. In 1928, however, he described newly discovered fossils in the Adelaide series, which he doubtfully referred to the so-called Lipalian interval. This latter paper attracted a great deal of attention, since in it he figured a possible pre-Cambrian eurypterid.

The present memoir, which is a continuation of this earlier work, was made possible by a grant from the Royal Society. The money was expended in opening two quarries, one at Tea Tree Gully, 11 miles northeast of Adelaide, the other at Beaumont, a suburb of that city. About fifty-five tons of quartzite were obtained, of which some seven tons were meticulously split into thin slabs. The material described and figured by David in 1928 was so fragmental that even in Australia, where the specimens were available for inspection, skepticism as to their being authentic was widespread. The newly quarried material, however, yielded better specimens, and it is stated that practically all Australian scientists are now agreed that they are genuinely organic.

The fauna of the Adelaide series is found in three horizons, in order from the oldest to the youngest, (*a*) the Tea Tree Gully, (*b*) the Beaumont, and (*c*) Brighton. The Sturtian glacial beds intervene between the Brighton and the Beaumont formations. The Tea Tree Gully fauna contains giant annelids and arthropods, the largest of the latter having an estimated length of approximately 10 feet! The Beaumont fauna con-

tains giant annelids and arthropods, a possible ancestor of the cephalopods, a large phyllocarid, problematical minute brachiopods, and a probable pteropod. The annelids and arthropods of this fauna show a considerable evolutionary advance over those from the Tea Tree Gully fauna, which is stratigraphically about 3,500 feet below the Beaumont beds.

The Brighton assemblage is characterized by a marvelously preserved, apparently dwarfed fauna of annelids and arthropods, partly postglacial, partly glacial in age. The writers suggest that the severity of the Sturtian ice age accelerated, if it did not indeed cause, the almost complete extinction of the giant annelids and arthropods of the Tea Tree Gully and Beaumont horizons and that the Sturtian refrigeration was a catastrophe to the contemporaneous marine life similar to the one during the late Paleozoic glaciation.

There is a question as to whether the Adelaide series should be considered (1) an infra-Cambrian part of the Paleozoic, (2) a Lipalian sequence, or (3) a definite Proterozoic series. The writers, however, stress the following points: (*a*) The Tea Tree Gully beds are stratigraphically 17,000–23,000 feet, and the Beaumont 14,000–20,000 feet, below the definitely lower Cambrian Protolenoid horizon in the Archeocyathid limestone series. (*b*) At least one erosional interval is recorded in the Adelaide series. (c) The Sturtian tillite probably is to be correlated with the Keweenawan [*sic*] tillite of North America, the Numees tillite of South Africa, and the Yangtse tillite of South China. (*d*) The fossils themselves are all vastly different from any previously described Cambrian organisms. In fact, they differ more from the Cambrian types than do the Cambrian fossils differ from the Carboniferous. Accordingly, Professor David was of the opinion that, although the Brighton fauna might possibly be as young as Lipalian, the Tea Tree Gully and Beaumont faunas lived during the later Proterozoic.

In the present memoir only five fragmentary specimens from the Tea Tree Gully horizon are dealt with. These are two closely related species belonging to the new genus *Protadelaidea*, of which the genotype is *P. howchini*. Professor Tillyard considered these species the sole known representatives of a new class, Arthrocephala, of the phylum Arthropoda. This new class is characterized by the four unfused segments which form the prosoma, or head region, and by the primitive structures of the abdomen, which does not carry segmented appendages. Professor Tillyard has had the temerity to include relatively detailed restorations of *Protadelaidea* showing both the dorsal and the ventral surfaces. He has care-

fully compared *Protadelaidea* with a number of other arthropoda, such as the Onychophora, the Crustacea, and the Trilobita. The conclusion is reached, however, that it was arachnoidian in type and had a common ancestor with the eurypterids. A significant relationship is also noted between the Arthrocephala and the Crustacea indicating a common ancestry for these two large classes earlier in the Proterozoic.

The present publication was to have been merely the first of a long series describing the other pre-Cambrian finds, but since Professor Tillyard met death by automobile accident on January 13, 1937, the work undoubtedly will be greatly delayed. It is to be hoped, however, that someone can be found who will be able to carry on these extraordinary researches of David and Tillyard with the same high enthusiasm with which they have so long prosecuted the work.

Notwithstanding the great reputation of the authors, some critics are likely to wonder how it was possible to make the restorations on the basis of the type of specimens illustrated in Plates I–VI; and this wonder may persist despite the fact that many of the Australian scientists who have examined this material are inclined to say that "seeing is believing." Furthermore, few North American geologists are likely to agree that

> only in a country like Australia, one of the most rigid shields of the earth, where even lower Cambrian strata retain in places their original horizontality, and the rocks are only slightly compacted, and where fossils like trilobites are exquisitely preserved, can one expect to find even the Proterozoic rocks so little altered as to preserve traces of former fossils.

CAREY CRONEIS

Atomic Structure of Minerals. By W. L. BRAGG. Ithaca: Cornell University Press, 1937. Pp. 292+xiii; figs. 135; pls. 9; tables 12. $3.75.

Three chapters covering the first thirty-eight pages of this book are devoted to the geometry of crystal structures, the methods of X-ray analysis, and the principles of mineral chemistry. Mineral structures are described in the remaining thirteen chapters, the last eight of which cover silicates in 134 pages, leaving ninety-six for all other minerals. A table of space-group nomenclature and indexes complete the work.

Those wanting specific quantitative data on minerals may well turn to the *Strukturbericht*, and much more complete works are available covering the ground of the first three chapters; for example, Volume I of *The Crystalline State* by the same author. Nevertheless this book should prove of great value for students of mineralogy at the college level. The de-

scriptions are in general interesting, even surprising in their unevenness—thus six pages are devoted to retelling the Braggian explanation of the relation of indices of refraction to structure in calcite and aragonite. The statement (p. 37) that "minerals are essentially extremely stable, they would not otherwise be minerals" is an unfortunate one, as is the one (on p. 8) that "the choice of crystallographical axes and standard planes is arbitrary," though this is later rectified (p. 24). Mineralogists owe the author and the Baker Foundation thanks for bringing this scattered information into such handy compass; sooner or later most of it will be found in mineralogical texts.

D. JEROME FISHER

Geomorphologie der feuchten Tropen. By KARL SAPPER. ("Geographische Schriften," Heft 7.) Leipzig und Berlin: G. B. Teubner, 1935. Pp. 154; pls. 4; figs. 7.

It is generally recognized that during most of the earth's past the climate has been considerably milder than at present, and the tropical belt greatly extended, yet the tropical physiographic processes and their effects on sedimentation have hitherto received little systematic investigation. Now, however, Professor Karl Sapper, who has studied the Tropics, especially Central America, for the past forty years, has made use of his own observations as well as those of other German explorers in preparing this work, which is of great importance to geologists for its indirect bearing on historical geology as well as for its considerable additions to physiographic knowledge. There is, for example, his recognition that a very coarse, unweathered arkosic and conglomeratic deposit, such as has often been regarded as indicative of aridity, is customarily found near a highland in the moist Tropics, where the streams have become very deeply incised in bedrock because the luxuriant vegetation protects the soil mantle and greatly retards lateral erosion. Sapper discusses the influence of climate on the various physiographic processes and the consequent effects of the latter on the morphology of the land. He divides the Tropics into the continuously humid, the seasonally humid, and the cold Tropics, discussing at length both erosion and deposition in each. Three chapters are devoted to the effects on man, animals, and plants, and on the sea.

An elaborate discussion of this important work appears in the *American Journal of Science*, Vol. XXXII, pp. 297–306, and should be consulted by geologists who cannot read readily the German volume.

F. C. MACKNIGHT

"Tavole fondamentali per la riduzione dei valori osservati della gravità." By G. CASSINIS, P. DORE, and S. BALLARIN. N.S.:N. 13. Pavia: R. Commissione geodetica italiana publicazioni, 1937. Pp. 119+xxvii.

These tables are, as the title indicates, of geodetic origin, and are in fact essentially the same as the multigraphed tables distributed in 1933 at the Lisbon meeting of the International Association of Geodesy. The original brief Introduction in French has been replaced by a much longer one in English and Italian in parallel columns. Some auxiliary tables have been added and the tables of densities of compensation that appeared in the 1933 multigraphed edition have been suppressed for reasons hereafter suggested.

They are called to the attention of geologists interested in the study of subsurface inequalities in density by gravimetric methods, because they appear to be the only adequate tables of their kind adapted to any assumption as to density: complete isostatic compensation of the Pratt or Airy type within the limit of the tables (200 km. depth), incomplete compensation, no compensation at all, or any combination of circumstances that the mind of the geologist may devise. Hayford's tables are for a typical density of surface rock (2.67) and for complete compensation of the Pratt type at a depth of 113.7 km.

The density factor is easy to supply, the Italian tables being for unit density, and the density tables of the multigraphed edition are perhaps omitted in the present one in order to tempt the computer to try out various possibilities.

The need of greater generality than is afforded by the purely isostatic tables is indicated by the fact that the Survey of India finds that small-scale isostasy does not apply to India, although the first statement of the idea of isostasy that received general attention was made by Pratt on the basis of the surveys of India then available; also by the fact that Vening Meinesz has found large oceanic areas where isostasy seemingly does not fit, such as areas of negative anomaly off island arcs and areas of positive anomaly in mid-ocean.

Hayford's tables are easy to use, but the Italian computers maintain that there is not enough difference between the general tables here reviewed and tables of the Hayford type to make it worth while to use the latter. Isostatic reductions in Italy have therefore been made with the multigraphed predecessors of the present printed tables. If special tables of a particular type are desired, they may be computed from these tables. In fact, the United States Coast and Geodetic Survey has used the multi-

graphed edition to revise and extend Hayford's tables and to change the depth of compensation.

These tables give what may be called the "direct effect," the vertical component of the attraction of a given mass. The horizontal boundaries of the masses, defined by the Hayford system of zones, are the same as in the Hayford tables. There is an "indirect effect" also to be considered. To make values of gravity at different places comparable they are reduced to the geoid, but the question arises: What geoid? Any supposition as to the density of subcrustal matter affects the geoid, thus giving rise to the indirect effect. Tables for computing the indirect effect wholly analogous to those here reviewed are given in *Special Publication No. 199* of the United States Coast and Geodetic Survey, "Tables for Determining the Form of the Geoid and Its Indirect Effect on Gravity." There is a gap in the tables in the region where they would be needed for computations according to the Airy hypothesis, but formulas and auxiliary quantities are given that facilitate filling this gap.

For complete compensation the indirect effect is usually small in comparison with the direct effect, but for the case of no compensation the indirect effect is four times the direct effect.

These valuable Italian tables are called to the attention of geologists interested in gravimetric work. Their use seems to be as simple as possible, but geologists should be under no illusions as to the amount of arithmetical work required.

WALTER D. LAMBERT

Economic Geology. By H. RIES. 7th ed. New York: John Wiley & Sons, 1937. Pp. 720; figs. 267; pls. 63. $5.00.

The seventh edition of Dr. Ries's *Economic Geology* is the latest of a long line that has served a most useful purpose throughout many years in the progress of economic geology in the United States. Discriminatingly compiled, well illustrated, and accompanied by selected bibliographies, they have served as reference works as well as texts. In the later editions increased attention has been advantageously devoted to metalliferous deposits and the principles of ore deposition. The seventh edition has been shortened slightly as compared with the preceding edition—but without weakening the presentation.

E. S. BASTIN

PUBLICATIONS RECEIVED

Aids to the Student of Conservation. Review and Thought-provoking Questions concerning Our Natural Resources and Their Conservation, edited by Parkins and Whitaker. By Dr. Stephen S. Visher. New York: John Wiley & Sons, 1937.

Annual Report of the Quebec Bureau of Mines for 1935. Part A, Mining Operations and Statistics. R. Paradis. 1936.

Bibliography and Index of West Virginia Geology and Natural Resources to July 1, 1937. By John B. Lucke. West Virginia Geological Survey Bulletin 4. Morgantown: West Virginia Geological Survey, 1937.

Bulletin de la Commission Géologique de Finlande No. 115. Helsinki: Imprimerie de L'Etat, 1936.

Bulletins 6, 8, 9, 10, and 11 of the Geological Section of the Newfoundland Department of Natural Resources. St. John's: Geological Survey of Newfoundland, 1937.

Comprehensive Index of the Publications of the American Association of Petroleum Geologists, 1917–1936. By Daisy Winifred Heath. Tulsa: American Association of Petroleum Geologists, 1937.

On the Displacement of the Tanna Fault since the Pleistocene. By Hisashi Kuno. Earthquake Research Institute, Tokyo Imperial University, 1936.

East Anglia and Adjoining Areas. By C. P. Chatwin. British Regional Geology. London: H.M. Stationery Office, 1937.

Geologische Kaart van Java. By R. W. Van Bemmelen. Toelichting bij blad 66 (Karangkobar). Batavia: Landsdrukkerij, 1937.

Gold Coast Geological Survey Bulletin 8. Colchester: Benham & Co., 1937.

History and Value of the Name "Catskill" in Geology. By G. H. Chadwick. New York State Museum Bulletin 307. Albany: University of the State of New York, 1936.

Illinois Mineral Industry in 1936: A Preliminary Statistical Summary and Economic Review. By Walter H. Voskuil, Alma R. Sweeney, and W. A. Newton. Division of the State Geological Survey, Report of Investigations No. 46. Urbana, Ill., 1937.

Ein Interglazialfund bei Rouhiala in Südostfinnland. By Gunnar Brander. Bulletin de la Commission Géologique de Finlande, No. 118. Helsinki: Suomen Geologinen Toimikunta, 1937.

Interrelationships of the Elasmobranchs with a Key to the Order Galea. By E. Grace White. Bulletin of the American Museum of Natural History. Vol. LXXIV, Art. II, pp. 25–138. New York, 1937.